# The negative relief of larger floodplains

Lewin, John; Ashworth, Philip J.

*Published in:*
Earth-Science Reviews

*DOI:*
10.1016/j.earscirev.2013.10.014

*Publication date:*
2014

*Citation for published version (APA):*
Lewin, J., & Ashworth, P. J. (2014). The negative relief of larger floodplains. *Earth-Science Reviews*, *129*, 1-23. https://doi.org/10.1016/j.earscirev.2013.10.014

Earth-Science Reviews 129 (2014) 1–23

Contents lists available at ScienceDirect

Earth-Science Reviews

journal homepage: www.elsevier.com/locate/earscirev

# The negative relief of large river floodplains

John Lewin [a], Philip J. Ashworth [b,*]

[a] Institute of Geography and Earth Sciences, Aberystwyth University, Llandinam Building, Penglais, Aberystwyth, SY23 3DB, UK
[b] Division of Geography and Geology, School of Environment and Technology, University of Brighton, Lewes Road, Brighton, BN2 4GJ, UK

ARTICLE INFO

*Article history:*
Received 13 June 2013
Accepted 28 October 2013
Available online 8 November 2013

*Keywords:*
Large rivers
Floodplain geomorphology
Topography
Sedimentation
River-floodplain connectivity

ABSTRACT

Large floodplains have multiple and complex negative relief assemblages in which depressions fall below local or general floodplain surfaces at a variety of scales. The generation and dynamics of negative relief along major alluvial corridors are described and compared. Such depressions are significant for the storage and passage of surface waters, the creation of a range of riparian, wetland, lacustrine and flowing-water habitats, and the long-term accumulation of organic materials.

Working on trunk channel remnants, drowned valleys and subsidence basins, fluvial processes modify floodplain negative relief through differential erosion and sedimentation. Effectively this takes place in three genetic zones: rheic, transitional and perirheic. We show that transitional zones marginal to active channels significantly diversify form complexes, and we demonstrate the diachronous nature of zonal processes and the complex nature and pace of depression modification and infilling. Four less well-understood sets of coupled phenomena are assessed: (i) floodplains associated with discontinuous river banks, (ii) the scales and types of scroll bar generation, (iii) factors underlying the contrasts between meander and braidplain surface relief, and (iv) the generation and function of large floodplain wetlands and lakes.

The survival likelihood of surface negative relief relates to geomorphological connectivity; this is described for each of the rheic, transitional and perirheic zones. The implications for contemporary aquatic system management are discussed. A key to understanding and managing negative relief on large river floodplains, and their associated ecologies and sedimentation, is to quantify *both* sedimentological and hydrological river-floodplain connectivity.



**Contents**



* Corresponding author.
*E-mail addresses:* john1lewin@btinternet.com (J. Lewin), p.ashworth@brighton.ac.uk (P.J. Ashworth).

0012-8252 Crown Copyright © 2013 Published by Elsevier B.V. Open access under CC BY license.
http://dx.doi.org/10.1016/j.earscirev.2013.10.014

## 1. Introduction

Large floodplains possess significant relief that may be picked out by standing or flowing water in wet environments (respectively 'lentic' and 'lotic' environments in ecology) (Fig. 1). Depressions may also be distinguished by wetland and other ecological communities. The negative components of this relief are multi-scale; they may form enclosed depressions from m to km across, or they may be morphologically-connected linear systems of a variety of shapes and sizes. In depth, they can be up to or exceeding the bank full depth of their major trunk rivers, with water levels that fluctuate with hydrological regimes.

In recent decades, studies of the relief and form dynamics of larger floodplains have been facilitated by the wide availability of remote sensing imagery, by the use of digital elevation models (DEMs) and by the improved dating of sediments. Collectively these show that larger alluvial floodplains have great complexity of form, dynamics and age (Dunne and Aalto, 2013). For example, Trigg et al. (2012) used Landsat Pan-Sharpened, 15 m resolution imagery to 'reveal' extensive networks of floodplain channels on the Amazon (Fig. 2A–B). The floodplain 'channels' are typically up to 40 m wide and 10 s of m deep, and function to distribute contemporary floodwaters and sediment (Fig. 2B), though there is often no simple linkage with the geomorphology of the

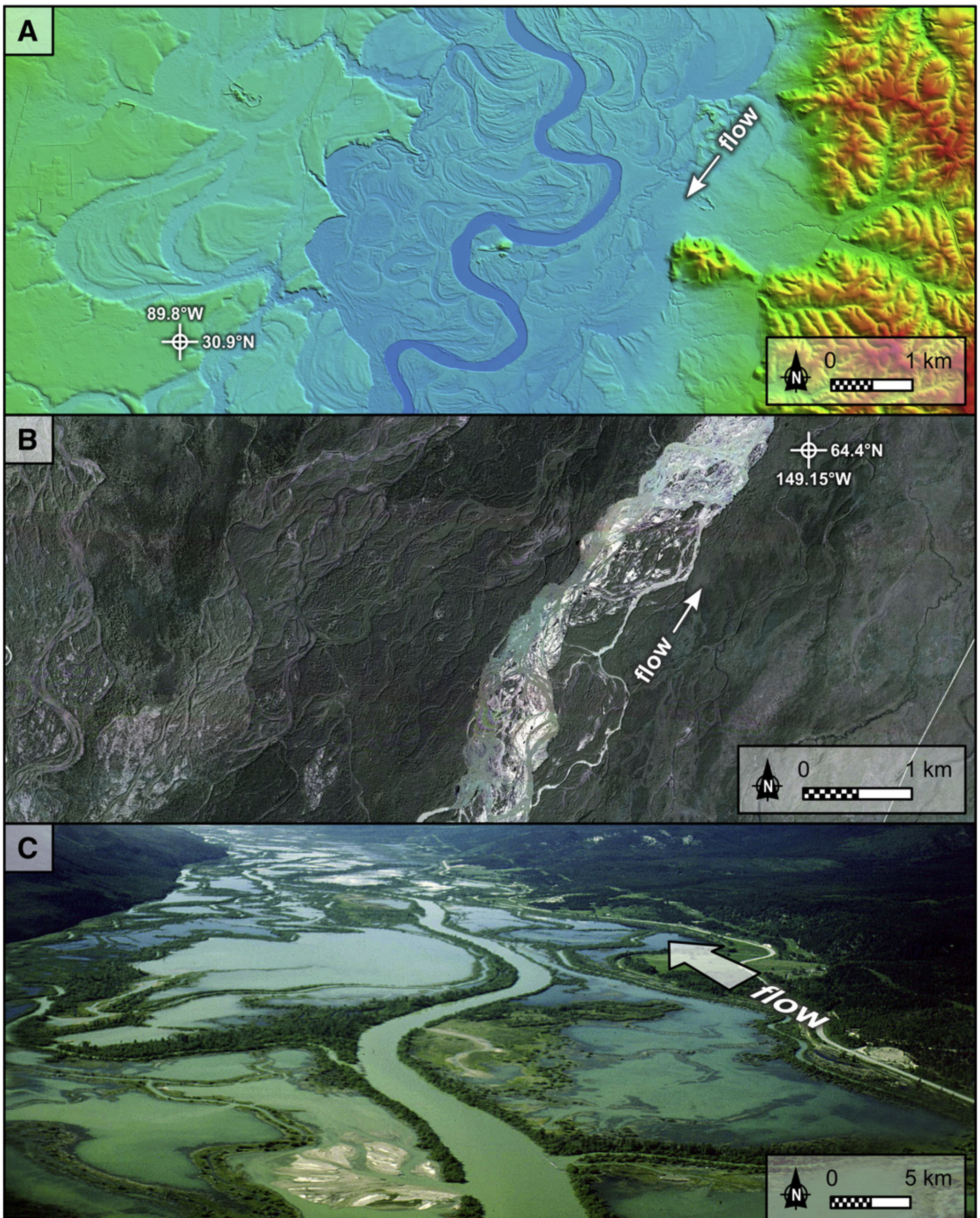


**Fig. 1.** Negative relief elements in floodplain environments: (A) meandering Pearl River, Louisiana/Mississippi, USA. Digital elevation model (USGS DEM) from NE quadrant of Bogalusa East quadrangle, Louisiana, UTM 15 NAD83, Louisiana FEMA project — phase 1: Florida parishes, under USACE (2001) contract [3008910ne.dem]; (B) braided Nenana river, Alaska, USA. Image taken in 2006. Image from USDA-FSA aerial photography field office, Salt Lake City, Utah. Orthophoto mosaic of Fairbanks North Star Borough, Alaska; (C) anastomosing Columbia River, near Harrogate, British Columbia, Canada. Image courtesy of Derald Smith.

contemporary major trunk stream (Lewin and Ashworth, 2013). In origin, the depressions they occupy may be swales on point bars, partially infilled palaeochannels, tributary channels, or internally-generated floodplain drainage nets (Fig. 2B).

The diversity and heterogeneity in relief and form of large floodplains mean that the pioneering floodplain classification based on mainstream channel association by Nanson and Croke (1992) is less effective for large floodplains, because they are (i) plural (reflecting the activities of several channels and sub-systems, with a partial disconnect with trunk stream activity), (ii) complex (with zonal differences in processing and rates of activity), and (iii) diachronous (with forms that have developed over a range of timescales that include the effects of Quaternary and even earlier allogenic systems).

This paper focuses on the fluvial geomorphology that generates larger floodplain negative relief elements, (Table 1) concentrating on valley corridors beyond the reach of coastal and tidal processes where non-fluvial processes also come into play (e.g., Fontana et al., 2008; Hijma and Cohen, 2011; Blum et al., 2013). We (i) describe and classify negative relief forms, big and small; (ii) explain how negative-relief forms are differentiated zonally by mode of generation, dimensions and survival history; (iii) contrast the nature of floodplain relief in low-bank, meandering and braided environments; (iv) consider geomorphological processes for channelled and non-channelled flow in larger waterbodies; (v) document negative relief survival times; and (vi) show why it is important to understand the evolution and filling of negative relief for restoration and habitat management. Large rivers are defined loosely as the ones with channels greater than 100 m wide (cf. Miall, 2006; Latrubesse, 2008; Ashworth and Lewin, 2012) and the forms examined here are meso- to macro-scale, from metres to kilometres across, within floodplains that are from c.10 to c.100 km in width. Their diachronous sub-aerial sedimentation surfaces incorporate boundaries ranging from Miall's 3rd to 7th order (Miall, 2006, his Table 4.2). Extensive lake areas (Section 2), fluvial/lacustrine interactions (Section 7) and channelized flood flows (Trigg et al., 2012) are

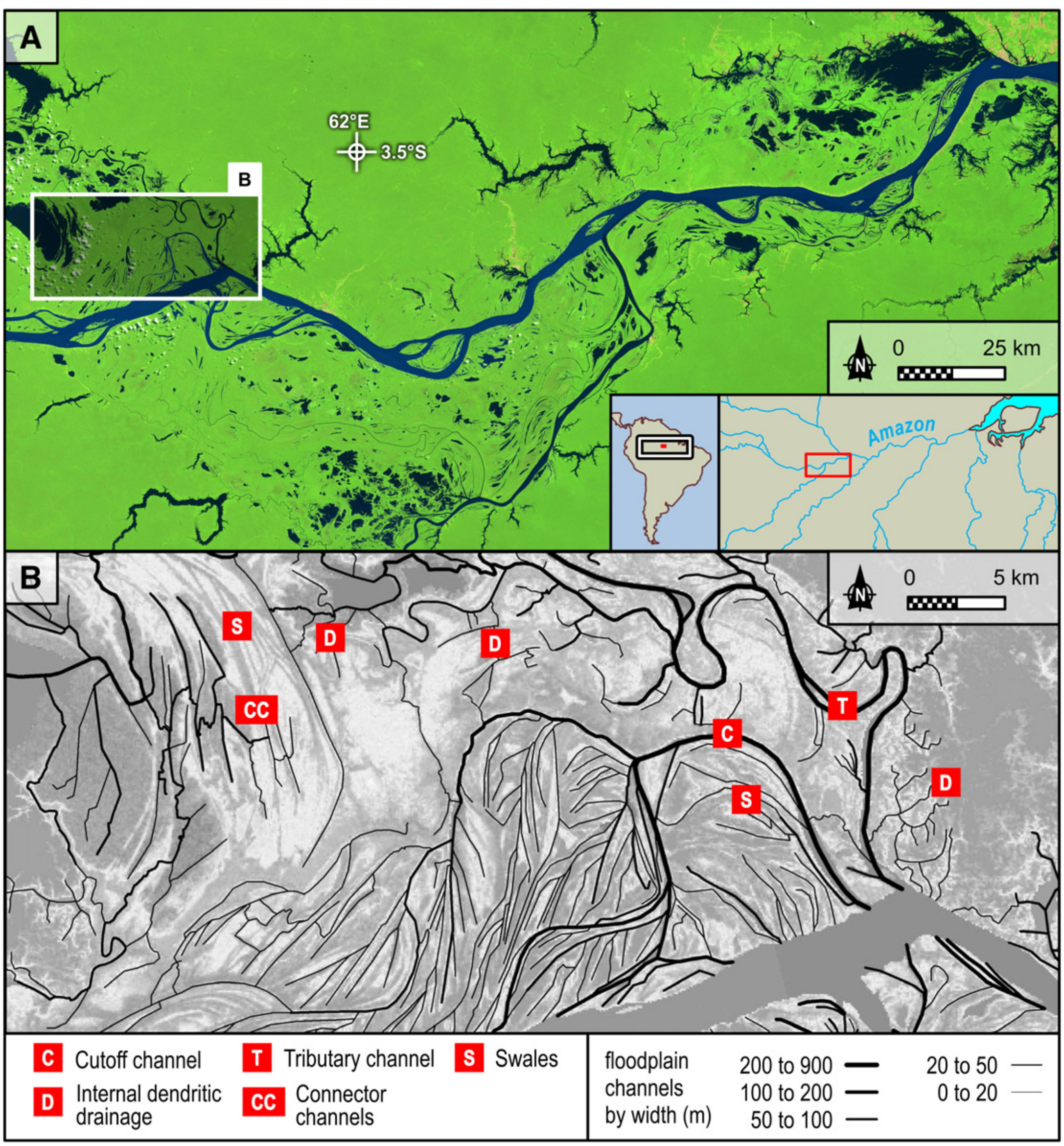


**Fig. 2.** (A) 300 km reach of the Middle Amazon, centred on the municipality of Anori, Brazil, that was mapped by Trigg et al. (2012); (B) floodplain 'channels' mapped by Trigg et al. (2012) for the inset box in (A) using Pan-Sharpened Landsat imagery at spatial resolution of 15 m and local fieldwork. Channels are depicted proportional to their water widths; some swales and depressions are not mapped as water-filled. Labelling of the different types of floodplain drainage added by present authors. Note that in this transitional zone many of the floodplain channels are hydrologically-connected to the trunk stream, but that they may have a variety of origins in terms of geomorphological development. Landsat imagery in (A) courtesy of the U.S. Geological Survey. Image taken on 11 August 2001.

**Table 1**
Meso- and macro-scale floodplain negative relief forms.

| Genetic zone | Negative relief forms | Examples |
|---|---|---|
| (i) Rheic | 1. Main channels (single or anabranching)<br>2. Accessory through-channels<br>3. Tributary channels (including dammed tributaries with outlet channels) | Mertes et al., 1996; Sambrook Smith et al., 2006; Gupta, 2007; Iriondo et al., 2007; Latrubesse, 2008; Makaske et al., 2009; Ashworth and Lewin, 2012; Rozo et al., 2012; Lewin and Ashworth, 2013 |
| (ii) Transitional | 4. Channel margin slackwater zones<br>5. Bar-shelter backwaters*<br>6. Contiguous channel remnants (semi-detached cutoff arms*)<br>7. Tie channels<br>8. Internal drainage channel networks | Lewin, 1983; Rowland et al., 2005; Day et al., 2008; Constantine et al., 2010; Grenfell et al., 2012; Toonen et al., 2012; Trigg et al., 2012; Dieras et al., 2013 |
| (iii) Perirheic | 9. Cutoff palaeochannel segments* (cf. 1, 2 & 3 above)<br>10. Accretionary swales and irregular unsedimented voids* (cf. 4–8 above)<br>11. Large-scale flood basins occluded by channel-belt aggradation* | Panin et al., 1999; Paira and Drago, 2006; Werritty et al., 2006; Constantine and Dunne, 2009; Citterio and Piégay, 2009; Hoffmann et al., 2009; Sidorchuck et al., 2009; Gautier et al.,2010; Toonen et al., 2012 |

*occlusion-related.
Genetic zones.
(i) As components of geomorphologically active channel systems.
(ii) With hydrological and morphological connection to channels (one- or two-directional flows).
(iii) With morphological detachment from main channels (though with possible hydrological connection at high flows due to overbank inundation, tributary inflow, or groundwater rise) and limited geomorphological activity.

the highly significant responses to this relief along many large river systems.

Fluvial morphogenesis for large floodplains comes in several linked modes. First, there are features produced by active channel erosion (Section 3). Channel scour at first removes sediment, then as rivers relocate, accommodation space for sedimentation is opened up. Relict channel segments may nevertheless be left partially open for extended periods (Section 8). On large river floodplains, these may formerly have been mainstream single or anabranching channels, or accessory and tributary ones (Ashworth and Lewin, 2012). As Gardner and Ashmore (2011) have argued, the scour by laterally migrating channels also creates a lower 'minimum surface' determining later deposit thickness for active systems. Over geological time periods, it is this 'combing' or 'working' depth (Paola and Borgman, 1991) that helps determine the preservation of sediments in the rock record both in-channel and as palaeoforms on and below the adjacent floodplain (Best and Ashworth, 1997; Brown et al., 2013; Snedden, 2013).

Second, positive alluvial relief by sedimentation is created over a range of space and time scales (Table 2). Because the production and preservation of individual bedforms or sediment packages are spatially uneven (Bridge, 2003; Horn et al., 2012), depressions are left at their margins and these may become long-lasting components of floodplain surfaces (Sections 3–6). Arising from such positive 'sediment occlusion', negative forms may be on a small scale, as in the swales between scroll ridges on point bars, or large-scale, as in the ponding of backswamps or 'impeded floodplains' (Latrubesse and Franzinelli, 2002) behind encircling levees or elevated channel belts, by blockage through fan extension, or between channel-belt alluvial ridges and bedrock valley sides. Such sedimentation may be observed in the short term using direct measurement or repeat imagery, but a timescale of centuries and millennia is required for understanding major floodplains. Essentially 'autogenic' processes (ones that are ongoing and associated with incident river regimes) overlap with 'allogenic' ones (resulting from environmental change). Palaeoform fragments in the form of fans, channel belts, islands and low terraces derived from earlier morphogenetic systems may be preserved on valley floors so as still to outline, or be incorporated into, regularly inundated areas.

Thus a third major influence on inland valley floor forms, additional to the operation of current alluvial processes, concerns longer-term tectonics and prior geomorphological settings. Many large rivers have not currently infilled large-scale available sedimentation space in areas liable to contemporary flooding (e.g., Latrubesse and Franzinelli, 2005;

**Table 2**
Meso- and macro-scale floodplain positive relief forms.

| Genetic zone | Positive relief forms | Examples |
|---|---|---|
| (i) | 1. Mid-channel bars (longitudinal, transverse, composite)<br>2. Side and point bars (including channel benches) | Moody et al., 1999; Ashworth et al., 2000; Santos and Stevaux, 2000; Best et al., 2003; Lunt and Bridge, 2004; Rice et al., 2009; Hooke and Yorke, 2011; Vietz et al., 2011 |
| (ii) | 3. Levees<br>4. Lateral accretion ridges and attached bars<br>5. Bed-material splays and channel plugs<br>6. Islands (commonly composites of 1, 4 and 5. May remain detached or incorporated into the general floodplain surface)<br>7. Overbank sedimentation, commonly infilling, in part, (ii) and (iii) in Table 1<br>8. Source-bordering aeolian dunes | Nanson, 1980, 1981; Brierley, 1991; Asselman and Middelkoop, 1995; Ferguson and Brierley, 1999; Aalto et al., 2003, 2008; Adams et al., 2004; Slingerland and Smith, 2004; Kemp and Spooner, 2007; Gurnell et al., 2012; Rozo et al., 2012 |
| (iii) | 9. Older abandoned valley floor islands<br>10. Substrate-erosional remnant islands<br>11. Abandoned valley-floor sides and alluvial terraces<br>12. Substrate-erosional valley sides | Latrubesse and Franzinelli, 2005; Bridgland and Westaway, 2008; Erkens et al., 2009; Blum et al., 2013; Lewin and Ashworth, 2013 |

Genetic zones
(i) Actively forming and emergent in-channel.
(ii) Transitional zone sedimentation being incorporated into floodplain surfaces and vegetated islands.
(iii) Palaeoforms in perirheic zone and at valley sides.

Roy et al., 2011; Rozo et al., 2012). Depressions on large river floodplains may be associated with foreland basins (peripheral, retroarc) including foredeep and backbulge, graben, half-graben and other fault complexes. Other neotectonic activities involving faulting, downwarping, and local river incision are also significant within large valleys (Mertes and Dunne, 2007; Syvitski et al., 2012). Quaternary valley incision linked to sea level fluctuations has affected large rivers like the Amazon, in that case for some 2000 km inland. These have all provided opportunities for surface water ponding with only partial infilling by fluvial sedimentation.

Finally, there are ongoing post-formational activities. Overbank floodwater passage and groundwater drainage may dissect alluvial surfaces and connect prior depressions, so creating and maintaining internal dendritic drainage networks. Shoreline progradation, lateral encroachment of vegetation, and basal sedimentation operate to infill lakes. Covering all of these, fine sediment infilling and capping may eventually create a more planar floodplain. However, negative relief survives because infilling is neither immediate nor complete. This may be because erosion maintains internal drainage nets, or because sediment dispersal, by water or wind, has been inadequate to eradicate negative relief forms.

Floodplains have commonly been treated as alluvial sedimentation phenomena. Considerable progress has been made in quantifying the processes of floodplain accretion and sediment exchange (e.g. Pizzuto, 1987; Walling et al., 1996; Dietrich et al., 1999; Aalto et al., 2008; Day et al., 2008; Erkens et al., 2011; Van Dijk et al., 2012; Dieras et al., 2013; Khan et al., 2013; Nicholas, 2013). Much less attention has been directed to understanding the origins, types and roles of floodplain negative relief and the impact this then has on sediment and organic sequestration and recycling (Aufdenkamp et al., 2011; Bastviken et al., 2011).

Accordingly, this paper provides a different perspective: where has mineral sedimentation *not* been fully effective in levelling-up floodplain surfaces? The extended survival of negative-relief elements is important to different disciplines. At the event scale, it is of hydraulic significance for the guided passage of floodwaters, and for standing waterbody extent and dynamics (Trigg et al., 2013). These relate to impoundment; to replenishment by direct precipitation, overbank flows and groundwater rise; and to drainage and infiltration. Second, over seasons and years, biological habitats are created and maintained here in complex and diverse mosaics of freshwater, wetland and riparian ecosystems (Tockner and Stanford, 2002; Hohensinner et al., 2011; van de Wolfshaar et al., 2011). This includes the generation of waterbodies that are involved in the transmission of diseases such as malaria (Smith et al., 2013). The different scales and relative heights of negative and positive relief, particularly on the very largest river floodplains, provide a range of sub-environments for different vegetation types to flourish and therefore to provide both floodplain stability and ecological diversity (Marchetti et al., 2013; Stevaux et al., 2013; Valente et al., 2013). Third, negative relief elements that last for decades, and indeed very much longer, may accumulate organic deposits. These provide archival niches preserving a palaeoenvironmental record (Gibbard and Lewin, 2002), they operate as terrestrial carbon sinks and, over the long-term, give rise to hydrocarbon resources (Thomas, 2013).

## 2. Prevalence of negative relief in large rivers

Some of the world's largest floodplains are characterised by extensive and water-filled negative-relief elements (e.g., Iriondo, 2004; Latrubesse et al., 2005; Ashworth and Lewin, 2012; Latrubesse, 2010). For example, 55% of the floodplains on the 33 large rivers studied by Syvitski et al. (2012; their Table 1) contained deep floodplain depression zones. These include major negative relief associated with the Mompox depression in the Magdalena River system (9° 10′N, 74° 33′ W), the Poyang depression and Lake Poyang (29° 07′N, 116° 17′E) and the Dong Ting at the intersection of the Yangtze and Xiangjiang rivers (28° 54′N, 112° 49′E). In South America, Sippel et al. (1992) reported that 11% of the mainstem Amazon floodplain in Brazil is covered by lakes ($n = 6510$) although only 10% of these lakes are larger than 2 $km^2$. In the Pantanal Basin in Brazil, accommodation space created by subsidence and increased humidity since the last glacial period has allowed the creation of an enormous seasonally flooded wetland dominated by alluvial sedimentation and is referred to as the world's largest continuous wetland (Alho et al., 1988; Assine and Silva, 2009; Latrubesse, 2010). On the Paraná, Paira and Drago (2007) report a figure of nearly 50% for permanent or temporary lentic (lacustrine) waterbodies. Remote sensing imagery shows that many are humic or 'blackwater' lakes with minimal sediment input even during flood stage. On the Amazon near Manaus, open-water floodplain tracts include meander swales, ponded tributaries, and extensive backswamps behind channel-belt sediments (Latrubesse and Franzinelli, 2002; Rozo et al., 2012). Hydraulic modelling of the Amazon flood wave along the main channel suggests that up to 30% of the main stem flow exchanges water with the floodplain (Richey et al., 1989). There are also small accessory sediment-rich channels (*paranás*) passing through the floodplain. These function especially at flood stage, when the system appears to operate as a giant elutriator allowing side channels to 'filter' water and finer sediment out onto the floodplain. These streams develop their own levees and may shift laterally (Mertes et al., 1996; Rozo et al., 2012). Alluvial forms may be dominated by shallow lateral-accretion *várzea* derived from such accessory channel migration rather than from mainstream activity (Paira and Drago, 2007).

Cold environments can also have extensive floodplain relief but this is not simply created by flood flows during channel migration, but instead is associated with inherited glacial landscapes or present-day paraglacial and periglacial processes (Fig. 3A–B). Extensive thaw ponds and thermokarst features dominate such tundra environments but often these are disconnected from the main river channel except during snowmelt floods.

Smaller floodplains with meandering rivers also have water-filled/wetland palaeochannels and backswamps, though in long-settled areas these may have been greatly modified because of sediment infilling following accelerated soil erosion (Lang et al., 2003; James, 2013), or direct floodplain engineering and construction (Lewin, 2010, 2013; Hohensinner et al., 2011).

The prevalence of water-filled negative relief elements is not just governed by the hydrological and geomorphological connectivity between the river main-stem and floodplain drainage (Lewin and Ashworth, 2013). As Mertes (1997) noted, in some of the world's largest rivers the inundation of floodplains can be spatially and temporally variable and not just caused by overbank flooding events. As Fig. 4A–C illustrate, if the floodplain is essentially dry before overbank flooding from the river channel occurs, as the water rises the zone of flooding expands parallel to the river as levees are overtopped along the main channel and floodplain channels (Mertes, 1997). However, if water is already present on the floodplain – due to groundwater (or 'hyporheic' water), flooding of local tributaries, runoff from surrounding slopes, direct precipitation or antecedent water from prior floods – then local water ponding can resist the incursion of water across the surface of the floodplain (Fig. 4F). Saturation of parts of the floodplain may occur prior to the crest of the flood wave and the onset of overbank flooding (Fig. 4D–E). Thus, some of the world's largest river floodplains are only partially inundated with water from main-stem rivers during floods (Mertes, 1997), and the local accumulation of floodplain water is not only governed by the morphology of floodplain relief but also the connectivity and interaction between surface and subsurface water. Subsurface hyporheic water connectivity, important both chemically and biologically (Castro and Hornberger, 1991; Mertes, 1997), is only considered tangentially here.

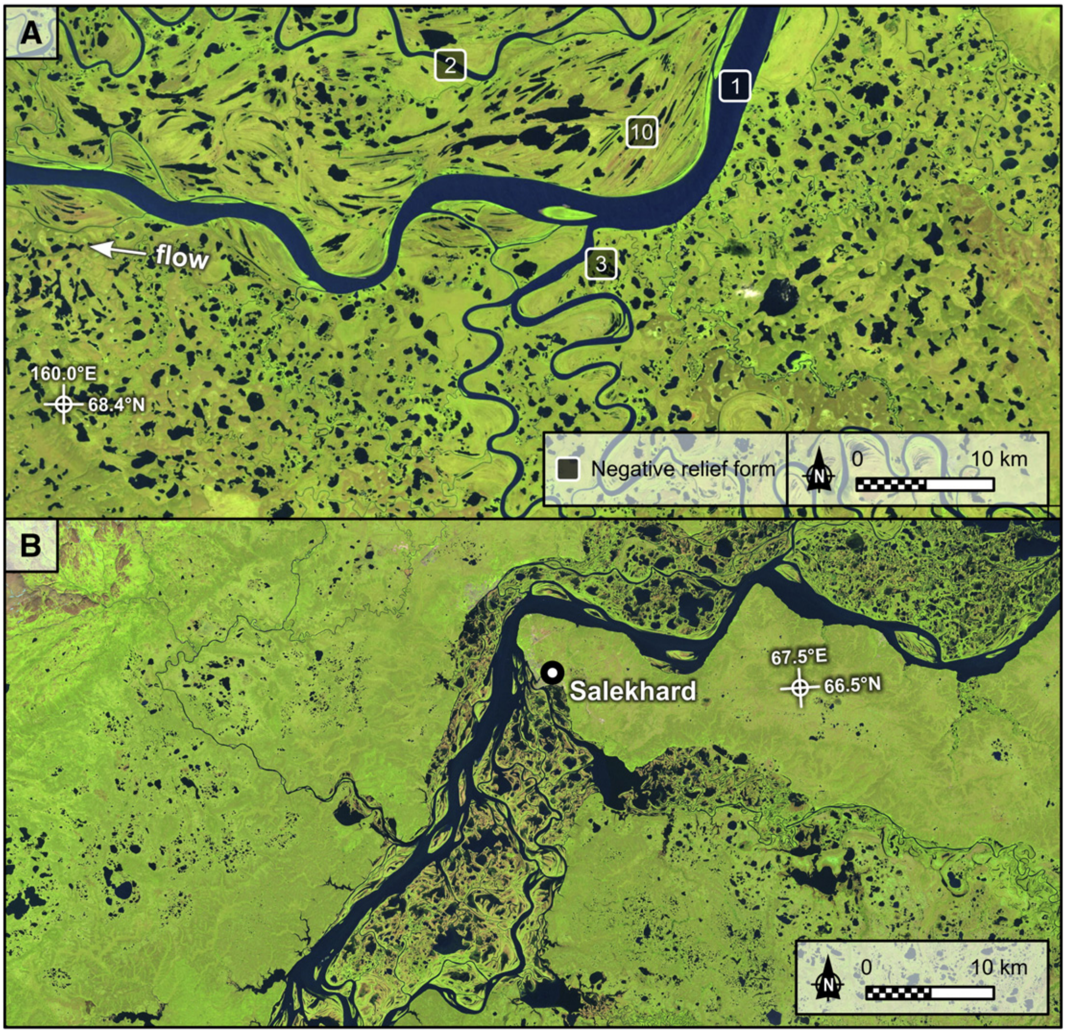


**Fig. 3.** (A) River Kolymar, Russia (68° 24′ N, 160° 00′ E). Extensive thaw ponds and meandering systems (both main, accessory and tributary channels) with some lateral accretion ridges. Image taken on 27 July 2002; (B) River Ob (66° 32′ N, 67° 30′ E) near its mouth at Salekhard in NE Russia. Large thaw ponds up to 4 km wide. Image taken on 23 June 2011. Landsat imagery courtesy of the U.S. Geological Survey. Label numbers in this and later figures are as shown in Tables 1 and 2.

## 3. Negative relief element generation

The range of *fluvial* processes contributing to negative relief creation (Table 1) may usefully be resolved into three genetic zones that differ in terms of process activities and form replacement rates:

(i) Channel domains that are geomorphologically active. This rheic zone (that is, dominated by flowing surface water) includes main, secondary and accessory channels that have varying degrees of geomorphological activity and styles of channel pattern evolution (Fig. 5A).

(ii) Geomorphologically transitional but hydrologically connected slackwater zones, backwaters and contiguous palaeochannel segments. These may or may not receive significant influxes of (generally finer) sediment to modify their forms (Fig. 5B).

(iii) Perirheic zones (Mertes, 1997) that are disconnected, both geomorphologically and in terms of hydrology, at all but extreme overbank flows or groundwater rise (Fig. 5C–D).

Most negative forms of fluvial origin are initiated in rheic domains (both main and other channels), they are transitionally but often considerably modified at rheic margins, and they may survive under slow infilling in perirheic environments. Boundaries may approximate those of abandoned channels, or may more closely be related to the margins of infilling sediment or occlusion bodies. The former reflect prior-channel dimensions such as width or radius of curvature. The latter are less likely to be simple elongate forms, and cover a considerable range of sizes, some marked by shelf-like inset fills, and lobate or digitate outlines. But the groups overlap: cutoff channels are plugged by occluding bed sediment (Constantine et al., 2010), and palaeochannels may be linked-up hydrologically with less regular between-bar depressions to give a beaded internal drainage network (van de Wolfshaar et al., 2011). Point bar scroll relief can also play a positive role in guiding chute development (Zinger et al., 2011) and the formation of 'bifurcate meander bends' (Grenfell et al., 2012), with swale irregularity itself having a feedback link to prior slow or rapid bar extension episodes. In addition to marginal sedimentation adjacent to channel flow, enclosure of negative relief areas takes place on a much larger scale in relation to valley-margin or tributary water-ponding beyond the reach of main channel shifting and to non-fluvial tectonic controls.

## 4. Negative relief in rivers with low or discontinuous banks

Where rivers are laterally mobile, receding cut-banks on one side may not be rapidly or evenly balanced by sedimentation on the other because (i) point bars tend to be built at a lower elevation than cutbanks, and (ii) point bars tend to be shorter than the eroding portion of cutbanks because of channel curvature (Lauer and Parker, 2008). If a floodplain is in equilibrium this local imbalance should be evened-out primarily by overbank deposition (Dunne et al., 1998; Aalto et al.,

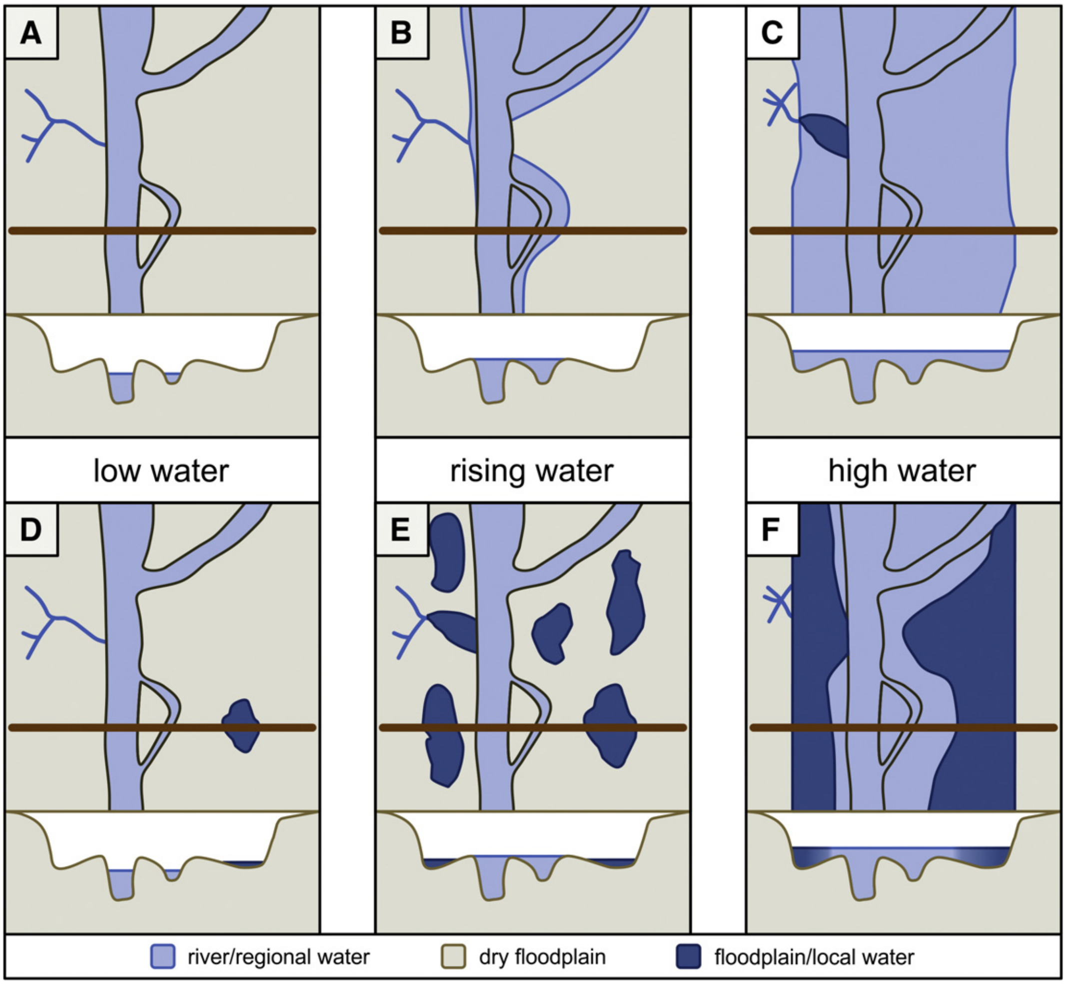


**Fig. 4.** Inundation patterns for a dry (A–C) and saturated (D–F) floodplain at three different water levels. The upper graphs in each box are planform views of a typical river with cross-sections shown immediately below with a vertical exaggeration on the order of 100. Mixing of water types are shown for surface waters only. Redrawn from Mertes (1997).

2008) and infill of negative relief elements such as abandoned stream courses or oxbow lakes. However, field studies show that diffuse sedimentation tends to be concentrated near the channel even on the largest river floodplains (Dietrich et al., 1999; Aalto et al., 2008).

On the non-eroding river flank, the bank can be ill-defined or locally absent. Fine sediment and buoyant organic material deposition takes place at channel margins in slacker waters and flow-detached zones where erosion produces a widening channelway that is not fully occupied by laterally-relocating river stemflow. In some freely evolving meander trains (Fig. 6), this may be: (i) against concave banks from which channelways are receding, (ii) in bar-tail 'shelter' zones, and (iii) at the downstream end of convex bends that are translating down-valley (Lewin, 1983). Where unfilled, voids may remain as open water, and eventually these get incorporated into floodplains. Alternatively, fine sediment influx may infill these voids to produce concave valley benches (Nanson and Page, 1983) or counter-point bar deposits (Smith et al., 2009). A detailed study of the hydrodynamics of deposition suggests that this in-channel deposition may be particularly effective at well-below bankfull stage (Vietz et al., 2011). Such thick, but spatially limited bodies of organic-rich sediment, are quite unlike fine sediments deposited overbank, and may also have different sediment and biological signatures to palaeochannel fills that are topped-up only during out-of-channel flows (Moody et al., 1999). Stratified fills (or a lack of them) depend on sediment supply, with flood units that may be coarser than in palaeochannel fills deposited further away from mainstream activity (Toonen et al., 2012). Under anthropogenically modified conditions, where sediments and pollutants are fed into streams at lower discharges, these within-channel marginal sediment stores may become much more important (Asselman and Middlekoop, 1995; Fryirs and Brierley, 2001).

Lower 'bank' elevations may also form entry points for inundating water to back-up and pass out into extra-channel areas. Conversely, and especially during stage lowering and floodwater return, these locations may have inset scour channels at exit points for floodplain drainage. These channels can be extended by headward growth to form internal floodplain drainage networks. Floodwater headcuts may in some circumstances be an alternative 'bottom-up' mechanism for avulsive channel relocation or chute-cutoff generation (Gay et al., 1998; Tooth et al., 2007; Zinger et al., 2011). To an extent, dryland rivers may also share some distinctive low-bank features, with aggradation and floodouts from intermittent flows, water transmission losses through percolation into alluvial deposits, and channel breakouts and splays (Tooth and Nanson, 2011; Tooth, 2013).

## 5. Negative relief created by scroll bar development

Fig. 6 shows a reach of the River Ob in western Siberia. Depositional scrolls on the inside of currently-occupied and abandoned bends create an undulating topography that provides abundant sites for establishment of km-long narrow water bodies. Deposition follows lateral relocation of the river by erosion of the bank opposite, and scroll irregularity is a lagged reflection of the spatial and temporal pattern of bank erosion upstream and on the other side of the river. This gives the accreting channel margin a ridged and tooth comb appearance

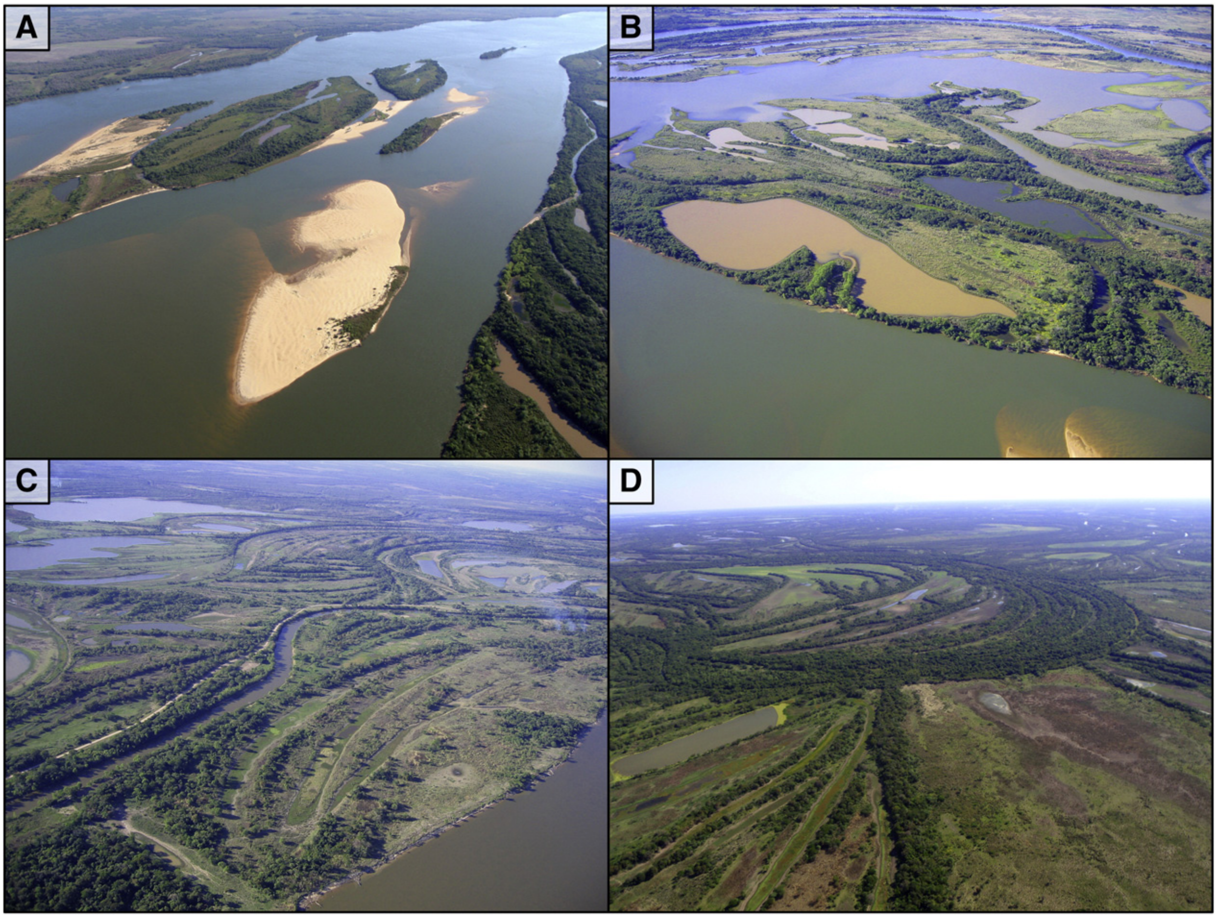


**Fig. 5.** Negative relief elements in the anabranching Rio Paraná, Argentina. (A) channel domains that are geomorphologically-connected and create elements such as topographic hollows, within-channel chutes and abandoned channels; (B) zones that are hydrologically-connected to the main channel and that may or may not receive influx of fines from suspended and washload; (C–D) perirheic zones that are mostly disconnected from the main channel and only fill during extreme floods or via sediment extrinsic to the fluvial system (see Table 3).

and there is a co-existing triple set of high-velocity main-channel stemflow, channel-side slackwater, and backwater arms. These have different degrees of intercommunication according to river stage. With low suspended sediment discharges but laterally mobile channels, there are greater opportunities for autochthonous organic deposition in irregular channel-side voids than in high suspended-sediment yield environments where the negative relief may be obliterated with fine sediment.

Point bar scrolls, despite their ubiquity and earlier pioneering research (Nanson, 1980, 1981), have attracted remarkably little

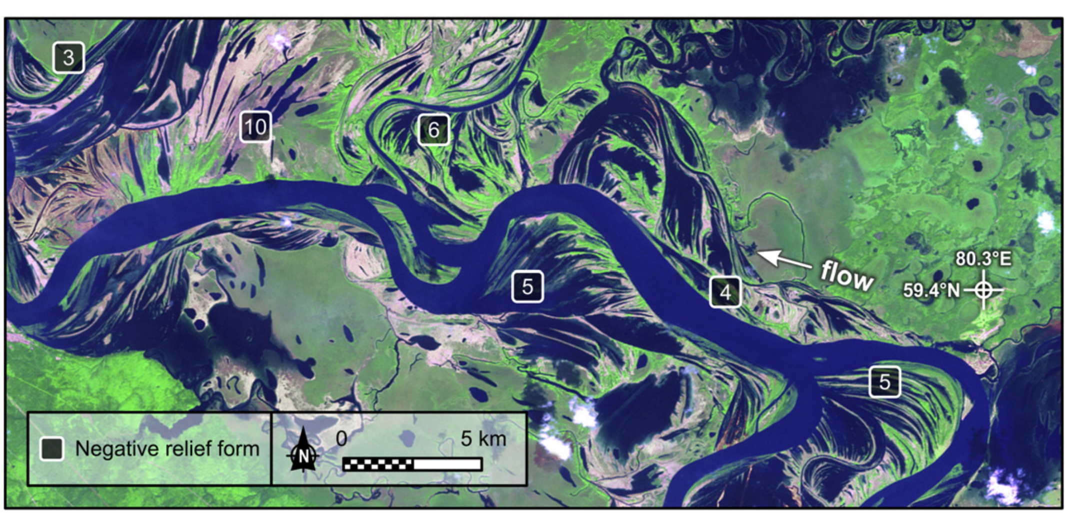


**Fig. 6.** Point bar morphology on the River Ob (59° 23′N, 80° 02′ E). Note cutoff channels, arcuate swales and more irregular water-filled depressions. Landsat imagery courtesy of the U.S. Geological Survey. Image taken on 30 May 2011.

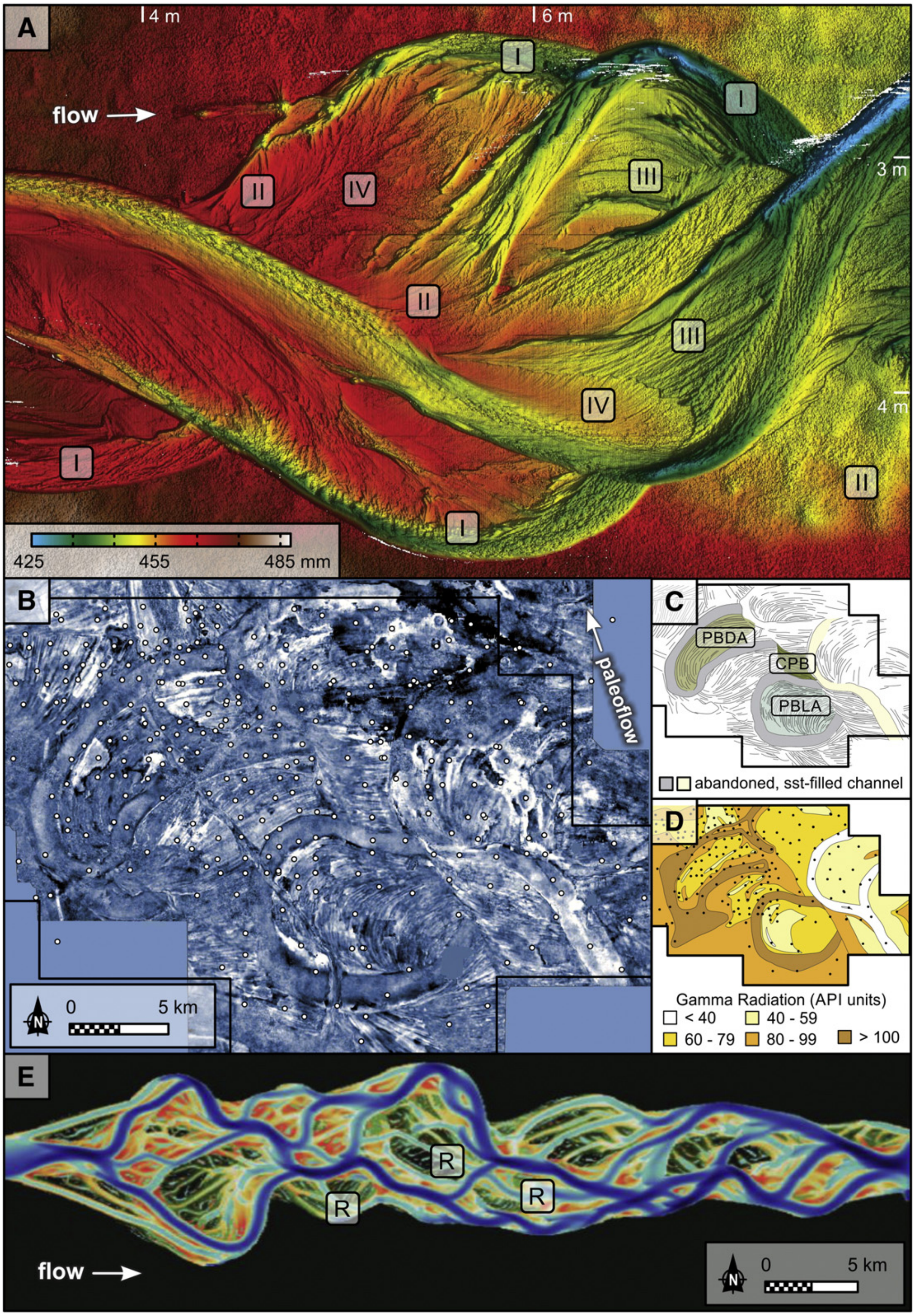


**Fig. 7.** (A) Digital elevation model of the central region of a meandering experimental model after 191 h of run time. The channel cuts through its deposits and disconnects former active channels (I) by a plug bar (II). Locations of scroll ridges and swales (III), as well as bar accretion on the point bar complex (IV) are shown. From Van Dijk et al. (2012, their Fig. 6); (B–D) (B) Seismic time slice taken through the strata of interest, 8 ms (~8 m) from a marine flooding surface present near the top of the McMurray Formation. The white dots represent well locations. (C) Line-drawing trace of the main features in B. Some of the main depositional elements are highlighted, including abandoned channel, point bar associated with lateral accretion (PBLA), point bar associated with downstream accretion (PBDA), counter point bar (CPB), and sandstone-filled channel. (D) Gamma-radiation map, constructed from values measured from wireline logs (wells used indicated with black dots) at the same stratigraphic interval of seismic time slice. The map provides a proxy for averaged lithology across the depositional elements observed in B and C. For example, the abandoned channels filled with siltstone are associated with the highest average gamma-radiation values observed. (B–D) from Hubbard et al. (2011, their Fig. 2, p.1126) with the depositional environment interpreted as dominantly fluvial with some tidal influence. (E) Example of a simulated channel morphology for a large anabranching river (from Nicholas, 2013; his Fig. 3G, p. 476). Image shows the morphology at low flow (10000 $m^3$ $s^{-1}$). Label R shows abandoned scroll morphology as discussed in the text.

attention as geomorphological elements contributing to positive and negative relief. Most detailed work has concentrated on understanding scroll bar evolution using either numerical models (e.g., Howard, 1984; Blondeaux and Seminara, 1985; Parker and Johannesson, 1989; Sun et al., 1996; Parker et al., 2011; Motta et al., 2012; Nicholas, 2013) or experimental work (e.g., Peakall et al., 2007; Dulal and Shimizu, 2010; Van Dijk et al., 2012) (See also Fig. 7A and E). Recent advances have also been made using high-resolution, three-dimensional seismic reflection data and drill cores that now provide unprecedented insights to ancient point bar and scroll morphology and sedimentology within hydrocarbon reservoirs that host some of the world's largest petroleum reserves (e.g., Smith et al., 2009; Hubbard et al., 2011; Fustic et al., 2012; Musial et al., 2012; Moreton et al., 2013; see Fig. 7B–D). Pioneering numerical modelling work on the morphodynamics of the world's largest anabranching rivers (Nicholas et al., 2012; Nicholas, 2013; Nicholas et al., 2013) suggests that scroll-bar morphology can dominate the early stages of pattern development (Nicholas, 2013; his Fig. 3B, D, and F, p. 476) and that large negative relief elements may remain after channel abandonment (labels R in Fig. 7E). In the field, negative relief elements may be linear, but larger semi-rounded polyhedral forms are also possible where a combination of 'jerky' autogenic channel change and event-driven channel shifts are followed by only partial infilling.

Individual ridges may prograde down-channel (Fig. 6) as a meander bend translates (Hooke and Yorke, 2011) or grow in arcuate form as the bend extends (Fig. 8A), with suspension material contributing to vertical growth (Rozo et al., 2012). Often, however, point bar scrolls accrete in a direction that is influenced by the local channel planform position at the time of maximum sediment transport, creating a mosaic of depositional packages (Fig. 8A), several scrolls wide and separated by an erosional-chute (labelled E in Fig. 8C) in between each patch of deposition. Individual scrolls on the Mississippi are up to 2 m high with erosional chutes 2–3 m deep (Fig. 8C). Elevated scrolls usually consist of finer traction or suspended load, and commonly have irregularly spaced negative-relief swales between them, though Fig. 8B–C show this is not always the case. Scroll relief is progressively smoothed away from the main channel as individual swales are filled with fine material (Fig. 8B). In planform, individual point bar scroll geometry is usually convex (e.g. Bridge et al., 1995; Leclerc and Hickin, 1997), but counter-point bars have scrolls that are concave and always locate down-river from or distal to the adjacent point bar (Smith et al., 2009).

Also on large and relatively straight rivers, compound sets of straight levee complexes may form as rivers oscillate from side to side and around islands (e.g., Rozo et al., 2012). These are rather different from lateral-accretion point bar scroll sets that are found on actively

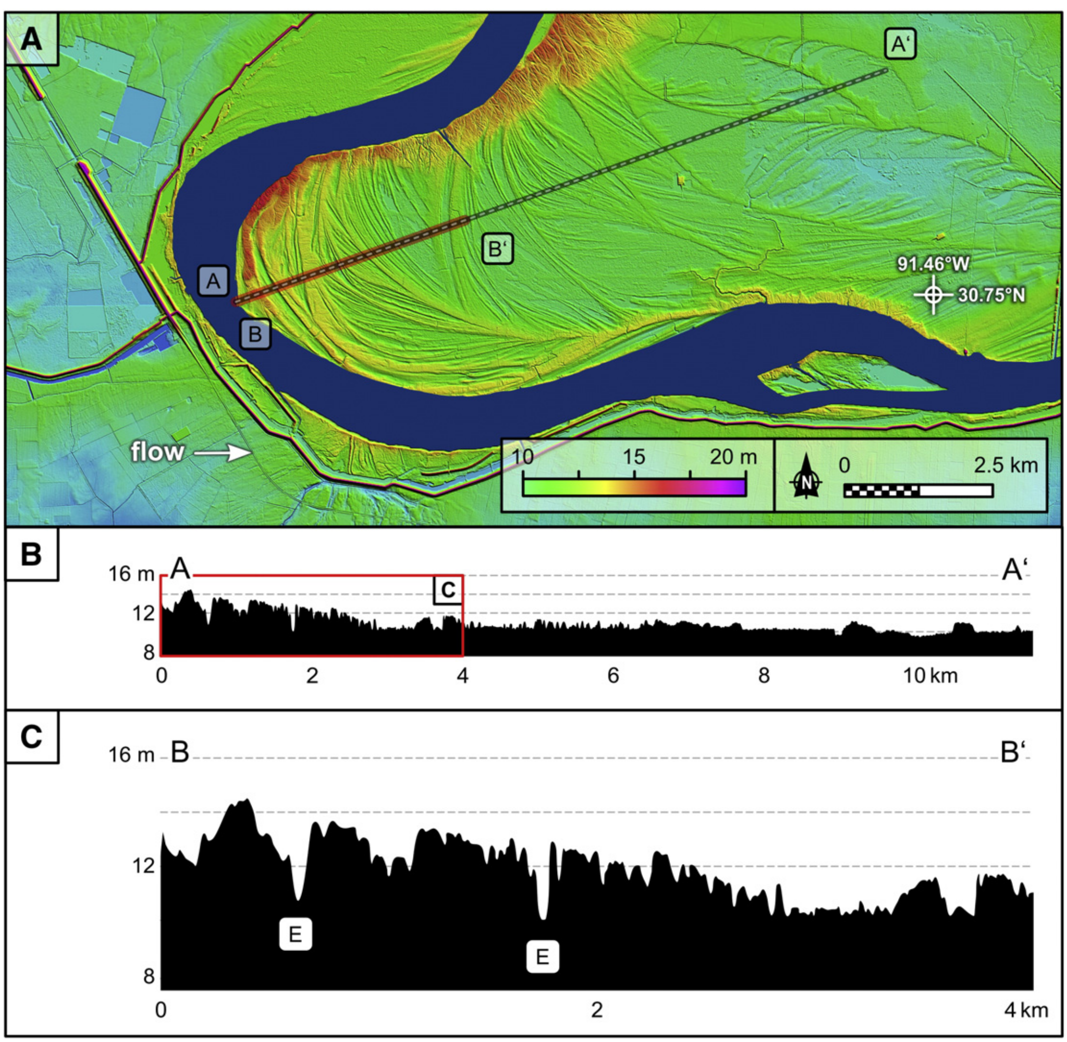


**Fig. 8.** (A) LiDAR image of scroll bar morphology on the Mississippi at Morganza, Louisiana, USA. Digital elevation model (USGS DEM) from Task Area 11 – North, Louisiana, Louisiana/Federal Emergency Management Agency (FEMA) project – phase 3 of Louisiana. Image taken in 2004. Note the patchwork of different accretionary packages related to different channel planform configurations and separated by up to 3 m deep chutes (labelled E); (B) cross-section through the scroll bar showing the gradual filling of negative relief (swales) further away from the main channel, and higher sedimentation levels near the contemporary channel; (C) detailed cross-section showing ridges up to 1 m high and mostly symmetrical, that are separated by narrower swales.

migrating secondary channels (Mertes et al., 1996; Latrubesse and Franzinelli, 2002). Short-term levee growth has been documented for larger rivers by Aalto et al. (2003, 2008). On large rivers transporting fine sediment, what may look like sets of accretionary bedforms may in fact be serial sets of extra-channel levees created as channels oscillate slightly in position over a timescale of many centuries.

In rivers with high suspended sediment loads, overbank and inner accretionary bank sedimentation, as well as bed-material point bars, closely track evolving planforms (Brooks, 2003; Page et al., 2003; Aalto and Nittrouer, 2012). Floodplain surfaces may be more planar, some with ridges developed in finer sediment, accentuating relief as floodwaters are channelled along swales depositing material preferentially on growing ridges between them. Coarser bed-sediment point bars may be weakly developed or absent. By contrast, other meander plains may not only have both cutoff and quasi-regular point bar swales tracking former channel migrations, but also more complex sets of irregular depressions left when combing depth pockets remain unfilled.

For many meandering rivers, cutoff channels may dominate floodplain relief (Figs. 1A, and 6) with their lengths related to parent-channel sinuosity (Constantine and Dunne, 2009). Water-filled or wetland arcuate forms may be coupled with relatively dry and smaller-scale point bar scrolls and swales (Fig. 9A–B). This applies to large tributaries of the Amazon, like the Purus and the Juruá, although, like the Amazon itself and its scroll-dominated plain, slight incision may mean that some alluvial forms occupy a low terrace (Latrubesse and Franzinelli, 2002; Latrubesse and Kalicki, 2002). The Purus high-sinuosity courses developed neck cutoffs with little counterpoint voiding, though they may also have secondary channels and long multi-arc abandoned reaches (e.g., at 4° 42′ S, 66° 45′ W). Meander bend cutoffs through erosion of a new channel across the neck of a bend ('chute cutoff') are less common in large rivers (cf. Lewis and Lewin, 1983), but when this does occur it can trigger the rapid delivery of sediment into the river at rates that are one to five orders of magnitude larger than those produced by lateral migration of individual bends (Zinger et al., 2011). Shifting confluence zones (Rice et al., 2008) may also locally create deep scours and what Brown et al. (2013) call 'soft avulsions' over extended time periods.

## 6. Meander and braidplain comparisons

In meandering rivers, palaeochannel dimensions may be maintained after cutoffs or avulsions occur. Curving forms can retain their recognisable bank outline on floodplains until filled by subsequent sedimentation, often over long time periods (Panin et al., 1999; Werritty


**Fig. 9.** (A) Meander morphology on the Rio Juruá (6° 31′ S, 68° 34′ W); cutoff domination, with ridges and swales. Image taken on 18 August 2011; (B) meander morphology on the Mississippi, USA (34° 39′ N, 90° 30′ W). Image taken on 2 November 2011. Landsat imagery courtesy of the U.S. Geological Survey.

et al., 2006; Citterio and Piégay, 2009; Kleinhans et al., 2011; Grenfell et al., 2012; Toonen et al., 2012). On braided rivers, the extra-channel equivalents of meandering system floodplains survive in the form of vegetated islands and surfaces (Reinfelds and Nanson, 1993; Thorne et al., 1993; Gurnell et al., 2009; Ashworth and Lewin, 2012; Ham and Church, 2012). This may arise because of channel relocation, avulsive abandonment, incision, or in the longer-term in relation to glacial/interglacial cycles. Replacement by less extensive and confined channel styles has occurred in mid-latitudes with Quaternary transitions from cold to warm stage alluvial systems that were less laterally extensive (Gibbard and Lewin, 2002; Lewin and Gibbard, 2010).

Braidplains exhibit more complex spatial and temporal relationships than in the distinctive three-stage sequence found in many meandering rivers (channel migration, cutoff and plugging, and overbank sediment infilling). Within active braided rivers, some channels are essentially geomorphologically inactive, although conveying water at high river stage (Ashmore, 1991; Ashmore et al., 2011). Accommodation space is set by deep channel scour (Best and Ashworth, 1997), with convergent and divergent flows creating evolving bedforms with island development, and new channels created by avulsive switching of main-channel flow (Kleinhans et al., 2013). As switching occurs and channel scour wanes, bed sediment may still be fed through diminishing branches and deposited there in sheets and lobes, obliterating the outlines of the once major channels (Lunt and Bridge, 2004; Rice and Church, 2010; Ashworth et al., 2011). For surviving alluvial surfaces, a critical difference between meander and braid plains is that channel abandonment has this longitudinally extensive waning, infilling and bed-material 'choking' process (Cheetham, 1979; Leddy et al.,1993). This is coupled with ongoing reworking of high-flow forms, and multi-channel abandonment results in a latticed surface of partially infilled palaeochannels and islands, rather than that produced by the three-stage cutoff one of meandering rivers (Toonen et al., 2012). Relict depressions may be narrow, linear and beaded along the line of a former channel, reflecting a history of occluding bed-sediment infilling. Relict channels may also preserve former confluence/diffluence zones giving Y-shaped branching forms with deep intersect scour pools (Kleinhans et al., 2013). Additionally, positive relief forms may also incorporate surface depressions, as in the case of infilled bar-top hollows or the stacking of infilled scours produced in the lee of large dunes (Best et al. 2006).

Fig. 10A–B show reaches of the braided Brahmaputra and Indus rivers. These have multiple high-flow channels and bedforms exposed at low flows, together with more permanent islands. The positive relief forms (Table 2) identifiable include mid-channel and side bars within active channels; lateral accretion ridges; and channel-choking infill. As channels shift, positive and negative forms evolve interactively within the zone of active bedload transport (Fig. 11). The braidplains of large rivers may be completely or substantially reworked over decadal time periods (see location F in Fig. 11D). Although only ~600 m wide,

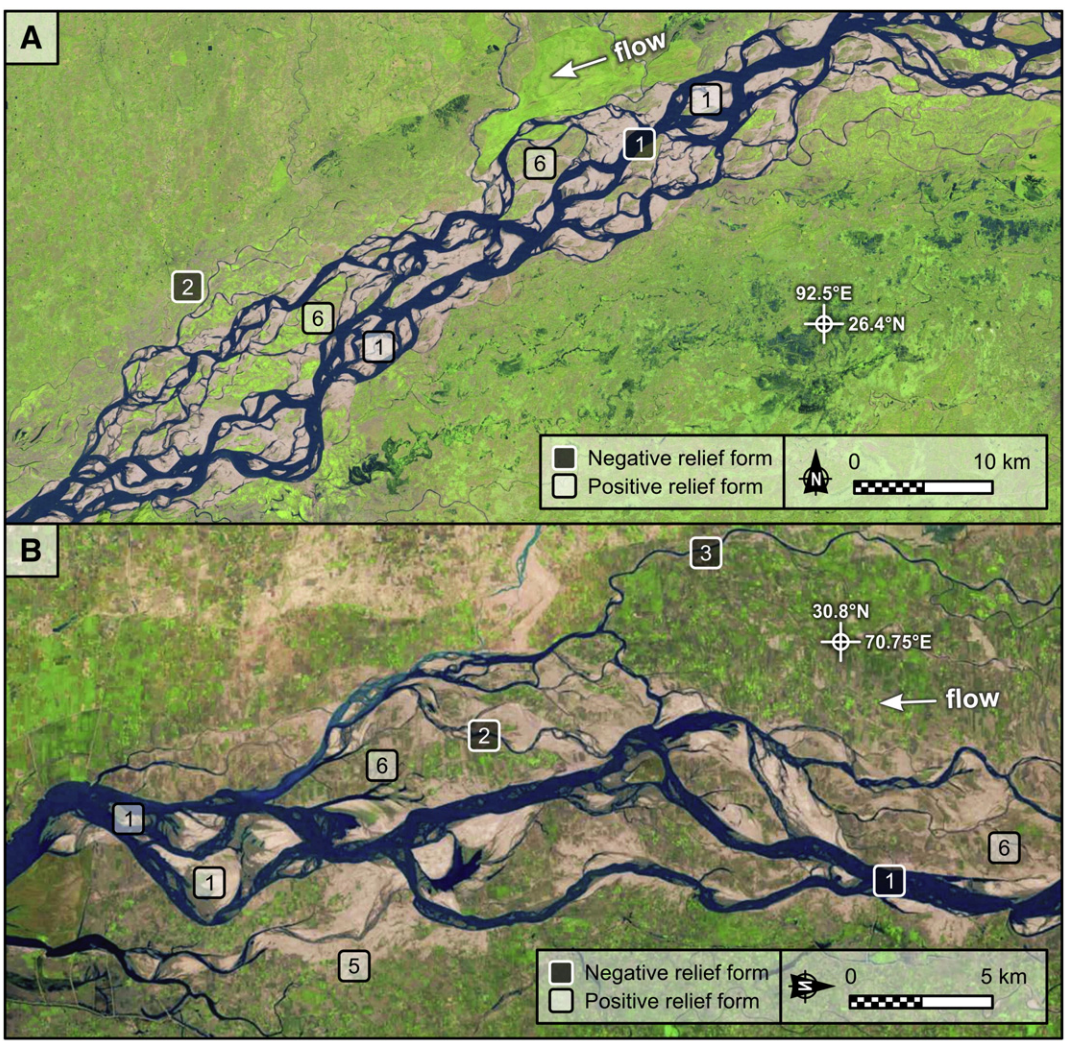


**Fig. 10.** (A) Braiding morphology of the Brahmaputra River, India (26° 21′ N, 92° 07′ E) showing vegetated islands and multiple dry channels (at this low-flow stage). Dimensions of former main channels are not preserved. Image taken on 10 November 2011; (B) braiding on the Indus River, Pakistan (30° 36′ N, 70° 50′ E). Large bar splays in active channel, and secondary 'choking' channels. Image taken on 18 December 2010. Landsat imagery courtesy of the U.S. Geological Survey.

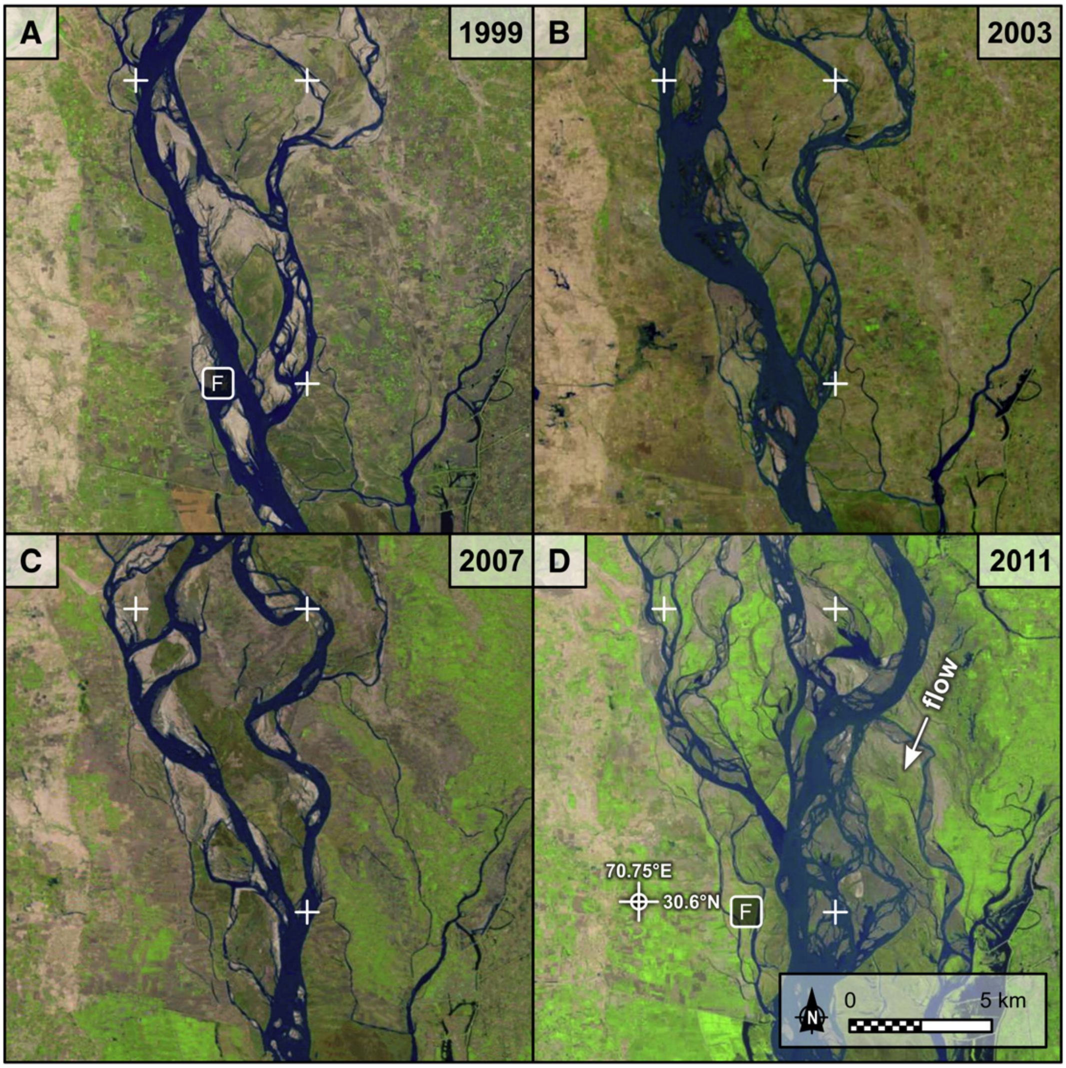


**Fig. 11.** (A)–(D) Indus River at four time periods showing rapid sedimentation/switching on a large river. Note that Fig. 11C has been edited to remove the bands of black line that occur on Landsat 7 images due to a failing sensor. Crosses on image allow location of channels through different time periods. Label F shows a zone of the braidplain that was active channel in 1999 but then abandoned, and vegetated floodplain in 2011. Images taken on (A) 26 November 1999, (B) 27 April 2003, (C) 31 October 2007, (D) 2 October 2011. Landsat imagery courtesy of the U.S. Geological Survey.

Sambrook Smith et al. (2010) showed that the braidplain of the sandy South Saskatchewan River can be completely reworked in a 1:40 year flood, and on the same river Lunt et al. (2013) used geophysical surveys to show that during 'normal' flood years the lengths and widths of individual bars may be truncated by up to 90% before preservation in the subsurface. Fig. 12 illustrates the rapidity of channel change that is possible in some sandy braided rivers, and hence the rate of turnover from positive to negative relief and vice-versa. Vegetated bars can remain relatively immobile and are trimmed at the margins (label X in Fig. 12A) but unit bars (label U in Fig. 12A–B) migrating through anabranches at rates of up to 1.5 m day$^{-1}$ are essentially transient features and usually only preserved as fragments in the subsurface (Parker et al., 2013). Up to 3 m deep scours can be created in the thalwegs and adjacent to resistant bank edges (label S in Fig. 12A–B) but these are filled rapidly by migrating dune bedforms (label D in Fig. 12C). Likewise, new scours can be created, abandoned and start to fill in less than a year (label A in Fig. 12C).

Thus the floodplain architecture resulting from braiding may actually have few width-preserved palaeochannel segments but rather depression 'strings' along their former alignments together with bedload fill outlines and drainage nets. The individual negative relief elements of braidplains are in general less pronounced, more complex in shape, and more frequent in number than in most meandering ones, though some of the latter (like the Ob, Fig. 6) have some of these characteristics as well.

Braided river process studies have commonly focused on 'early stage' processes: dune, bar and island development; bed scour and lateral accretion in active channels; and chutes, bifurcations and avulsions. Later-stage in-channel reworking and waning channel-infilling also contribute considerably to the relict topography of braidplains. *Relict* negative relief forms include isolated elongate scour holes (produced at former bends, bifurcating channels and channel convergence points) and extensive dendritic 'channel' talwegs, downstream of and between trimmed bar forms, and converging on the next set of scours (Doeschl et al., 2006; Lane, 2006; Kleinhans et al., 2013). The trimming and anabranching patterning of low flow channels may only in part bring out the complexity of relict lobate/dendrite forms created at high flows (Fig. 13). In section, braided river deposits may be dominated by lower-order inclined and horizontal strata sets and, unlike meandering systems, have few distinguishable bounding surfaces that conform to the dimensions of abandoned main channels (Bridge and Lunt, 2006; Ashworth et al., 2011; Lunt et al., 2013; Reesink et al., 2013). Quaternary braided-channel fills and surfaces may be equally lacking in major palaeochannel definition (Ashworth et al., 1999; Lewin and Gibbard, 2010).

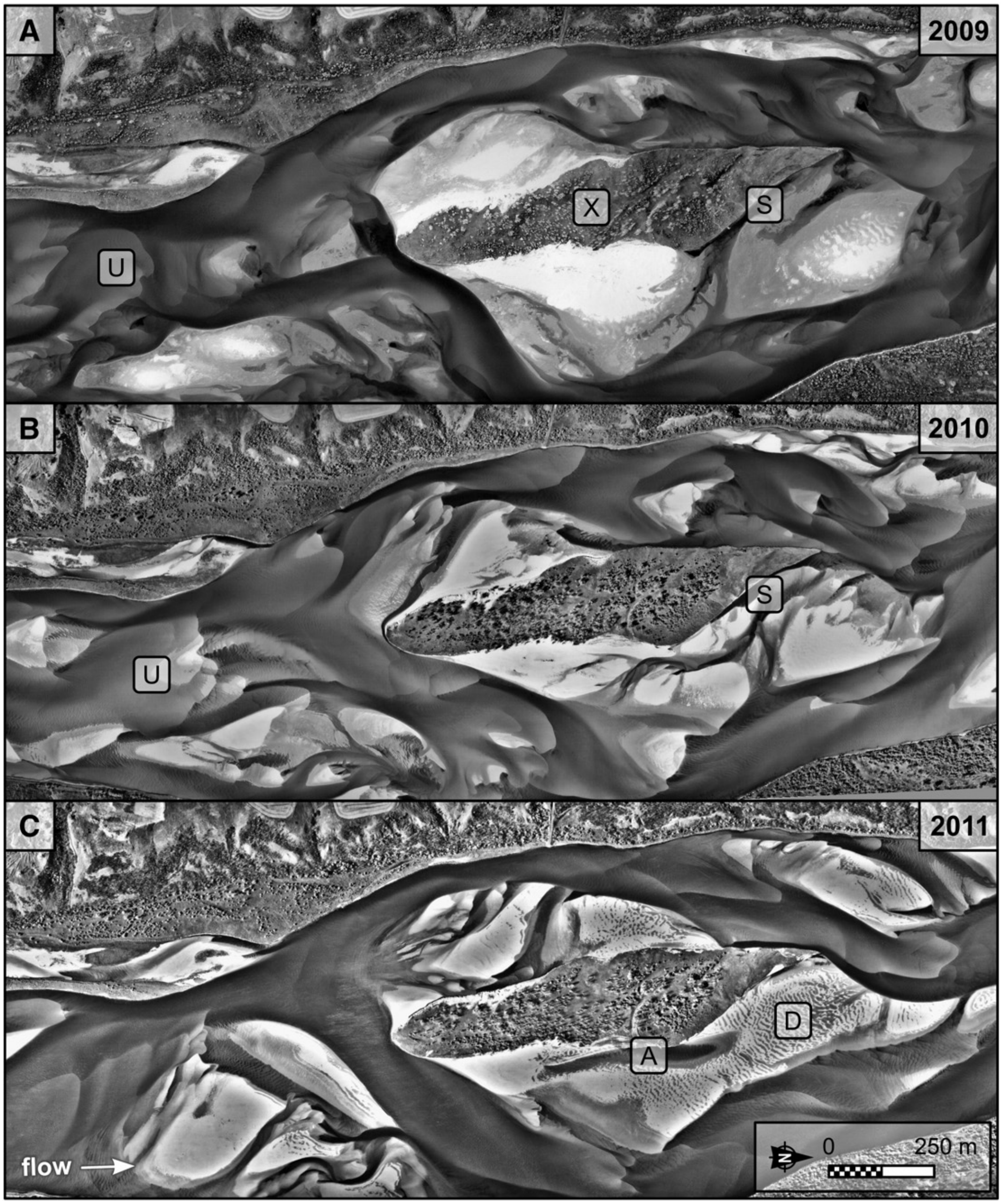


**Fig. 12.** Aerial photographs of the South Saskatchewan River, Canada taken on (A) 23 October 2009, (B) 18 September 2010, and (C) 11 October 2011. Note the rapidity of channel change and therefore the rate of turnover of negative and positive relief. Labels are described in the text.

## 7. Channelled and non-channelled water in major fluvial wetlands

According to Fraser and Keddy (2005), five of the ten largest wetlands in the world are riverine. These border the Amazon, Congo, Paraná (including the Pantanal), Mississippi and Nile. At the present time, other wetlands in cold-climate taiga and tundra zones, as in the Mackenzie River catchment or West Siberian Lowlands, are generally string bogs along minor rivers, or raised interfluve bogs. Some floodplain-located bogs effectively form positive relief forms there that have developed entirely through vegetative growth. Major river floodplains also have large bodies of seasonally open water in occlusion-impeded floodplain drainage zones, or dammed tributary valleys (Ashworth and Lewin, 2012; Latrubesse, 2010).

Many large tropical rivers replenish extensive associated waterbodies (Fig. 14A–C), together with seasonal flooding of forest and savannah bottomlands (Iriondo, 2004; Paira and Drago, 2006, 2007). This arises in particular because sedimentation has been linear and localized in channel belts (Fig. 14A–C) and inadequate to infill intra-cratonic basins and the wide linear depressions that major rivers occupy (Mertes et al., 1996; Archer, 2005; Latrubesse and Franzinelli, 2005; Iriondo et al., 2007; Valente and Latrubesse, 2011; Latrubesse et al., 2013). These include subsiding graben structures, valleys excavated in relation to lower relative sea levels, and tributaries back-ponded by major river aggradation.

Inter-channel waterbodies can also remain relatively sediment-free, despite the sediment-laden rivers flowing adjacent to them — even when there are open-water linkages between channels and ponded waters. With present-day large equatorial rivers, inundation waters may derive from runoff in lower parts of the catchments, and their levels may equilibrate with those in main channels that relate to floodwater inputs from distant headwaters. In practice there are complex adjustments of water level, but not a simple mainstream flood overflow involving dispersal of, and deposition from, sediment-rich waters

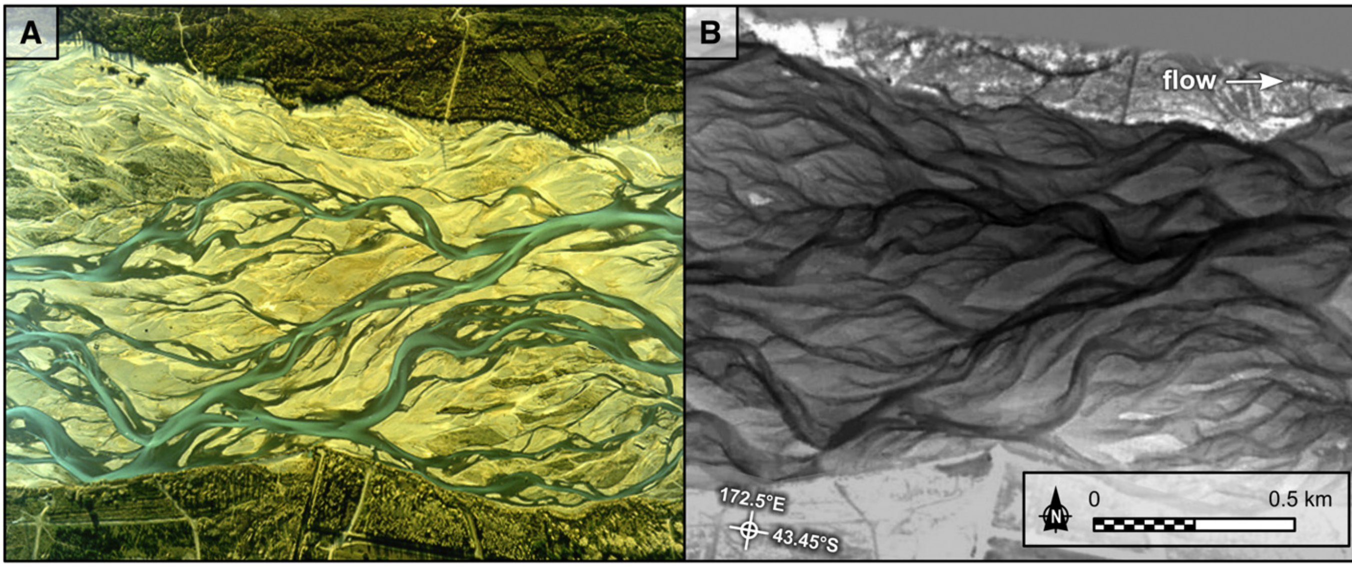


**Fig. 13.** (A) Air photo of the Waimakariri; (B) DEM of the Waimakariri River, New Zealand (43° 26′ S, 172° 29′ E). Note the relatively simple braided pattern of water-filled channels in (A), but the complex pattern of relief with widespread negative elements in (B). This includes attached dendritic elements as well as longitudinally connected channels. Images courtesy of Stuart Lane. Further details on the image processing technique are in Lane et al., (2003) and Westaway et al. (2003).

(Fig. 4B). In such circumstances, there is little advection of sediment from the main river; there is narrow diffusive contact between the juxtaposed waterbodies concerned, and a restricted tendency for crevassing. For avulsion it is the existence of a hydraulic gradient or improved flow efficiency following diversion that is required; typically this occurs under aggrading channel or channel-belt conditions (Törnqvist and Bridge, 2002; Slingerland and Smith, 2004). But if water levels in the extra-channel zone are similar to those in channels, then overflow, sediment dispersal and avulsion are less likely. A hydraulic barrier, as much as a bank of sediment, inhibits the lateral filling of broad valley troughs with sediment. Thus, subsiding environments like the wetlands of the lower Magdalena have quite narrow strings of mineral sediments within lacustrine fills (Smith, 1986). Such permanent waterbodies differ from those with markedly seasonal inundation deriving from mainstream overflow. Here advective transport may produce broad levees rather than narrow diffusive ones (Adams et al., 2004), whilst overbank deposition and avulsive relocation may distribute channel-belt sedimentation across wide valley floors (Valente and Latrubesse, 2011), decreasing the accommodation space for organic accumulation.

Marginal lakes and wetlands may have both channelled and non-channelled waters. On a small scale, an interesting but unappreciated process may be illustrated by the postglacial development of the Hay River that 'bridges' Lake Zama in Alberta, Canada (Fig. 15). In shallow water, the levee-lined trans-lacustrine channel has prograded to form a sinuous trail across the lake from upstream delta to exit point, together with some discontinued branches. These 'meanders' show very limited signs of lateral mobility, an activity commonly expected in the production of meandering planforms. But sinuosity here developed as part of initial downstream extension in a process quite unlike that of most meander models. Wetland river 'bridging' on a much larger scale can be seen on the Hwang He in China (29°54′N, 112°49′E), and in neotectonic basins on the Amazon (Latrubesse, 2010). Prograding levee arms are likely to be a major process for channel definition in shallow water across subsiding alluvial basins (where the rate of subsidence has been sufficient to 'drown' potential channel systems), or in the establishment of defined channels following avulsion into shallow inter-channel wetlands in anastomosing systems. Reviewing the establishment of new channels on floodplains following avulsion, Kleinhans et al. (2013) note the roles of incision or palaeochannel re-occupation, but neither may be the case for near-zero gradient lake/wetland channels.

'Rivers without banks' provide an important context both for contemporary aquatic biology and for the interpretation of ancient organic deposits. Differences between surface hydrological connectivity (with open water linkage), and geomorphological connectivity (involving sediment transfer), may allow negative water-filled relief to be long-lasting. With equilibrated water levels at high river stage, mobile biota can pass directly, and either way, between sediment-rich higher velocity channels and less-turbid backwater or lake environments through a kind of non-chemical osmotic transfer. This need not involve strong one-way overflow if extra-channel water levels are already high, though there may be subsequent falling-stage return flow when main channel levels drop. Such floodplain lacustrine environments can be remarkably sediment-free, long-lasting and without a chemically degrading influx of aerated water even where there are open-water connections. In the longer term, these conditions favour the production and preservation of freshwater organic deposits with little non-combustible content, as have many freshwater coal deposits (McCabe, 1984; McCabe and Parrish, 1992; Thomas, 2013). There may, however, be lesser nutrient input from main rivers than might otherwise be expected. Iriondo (2004) has pointed to distinctions between sand wetlands (with relict features from wet and dry climate phases and the reworking of dunes) and mud wetlands (with fluvial sediment input and greater nutrient availability), whilst Latrubesse (2010) has also underlined the importance of geomorphological factors in differentiating the nutrient and primary productivity status of lacustrine environments.

In non-coastal areas (where other factors including beach barriers are also important), there are some geological circumstances that appear favourable to large-scale terrestrial organic sedimentation of eventual economic significance: large rivers with sandy sediment loading draining through cratonic troughs (in wet climates) that they incompletely fill by mineral sediment deposition. The limited amount of non-combustible mineral content in many associated organic sediments suggests low lateral advective sediment transfer, even though there may be hydrological connection. Ponded waters already present hold back lateral dispersal from mainstream floods. Lower sediment loadings might also suggest minimal tectonic uplift and erosion in continental headwaters, but for the largest rivers their derived sediment may already have been deposited in foreland basins leaving cratonic lower courses relatively poor in terms of supplied sediment from cratonic tributaries (Ashworth and Lewin, 2012).

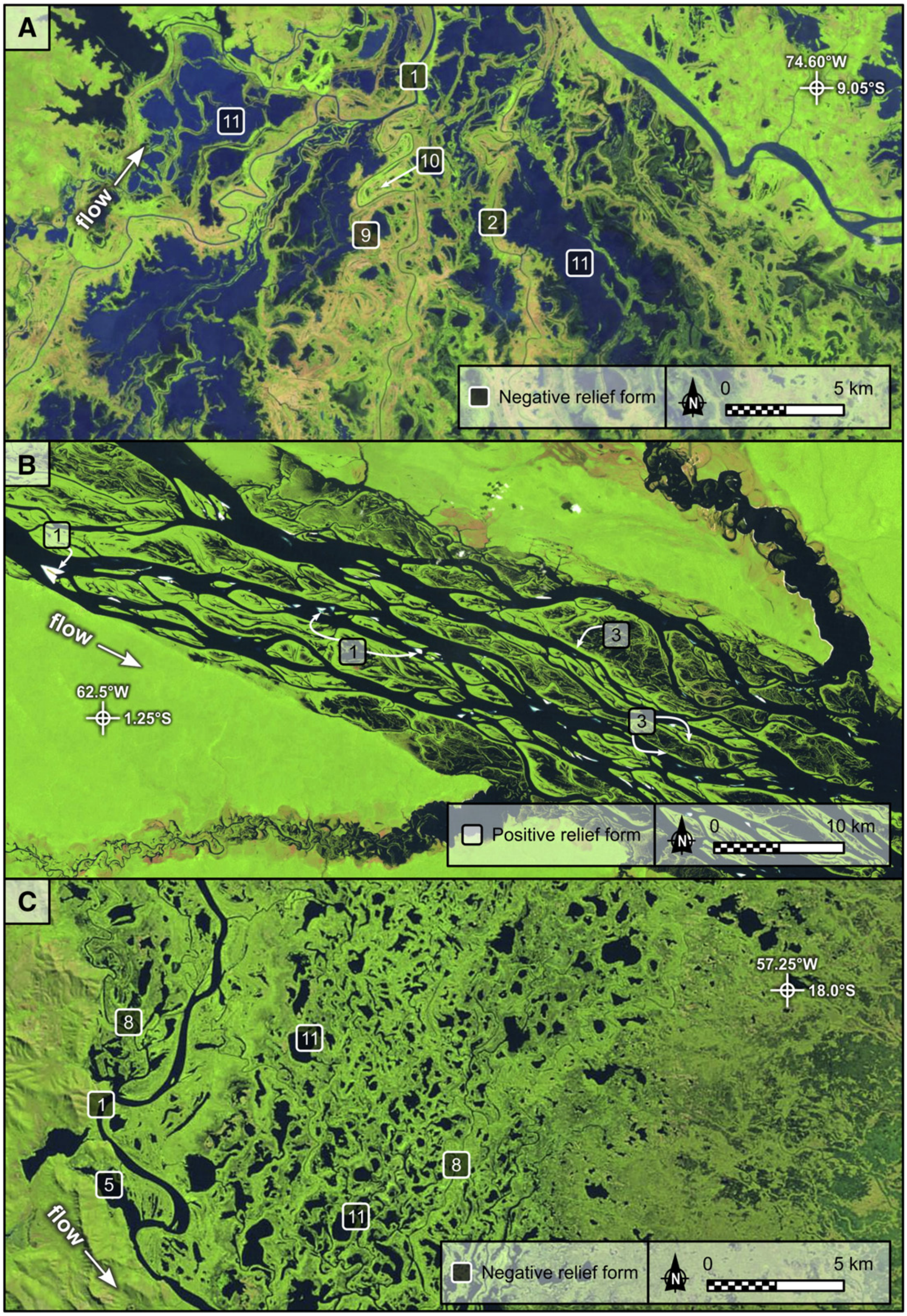


**Fig. 14.** (A) Anastomosing River Magdalena, Columbia (9° 00′ S, 74° 46′ W). See also Smith (1986). Multiple anastomosing channels with levees and splays into flood basins — some reaches laterally mobile. Image taken on 12 January 2001; (B) anastomosing on the Rio Negro, Brazil (1° 11′ S, 62° 17′ W) (cf. Latrubesse and Franzinelli, 2005). Note numerous islands, extensive levee systems with occluded lakes and wetlands behind. Some lateral accretion ridges. Image taken on 9 October 2011; (C) Paraguay River in the Pantanal wetlands, south-west Brazil (18° 05′ S, 57° 26′ W). (cf. Assine and Soares, 2004; Assine and Silva, 2009) showing an extensive and unconfined floodplain with swamps and lakes at the foot of the Taquari fan (cf. Makaske et al., 2012). Image taken on 24 May 2000. Landsat imagery courtesy of the U.S. Geological Survey.

## 8. Survival history of negative relief

Negative relief may be infilled via a range of fluvial and non-fluvial sources such as channel accretion by bedforms, sediment fall-out from overbank flow, authigenic mineral precipitation associated with flood and groundwater evaporation, organic growth (Tooth and Nanson, 2011), aeolian dust and sand input (Bullard and McTainsh, 2003), and, more rarely, volcanic and marine sedimentary input (Marriott and

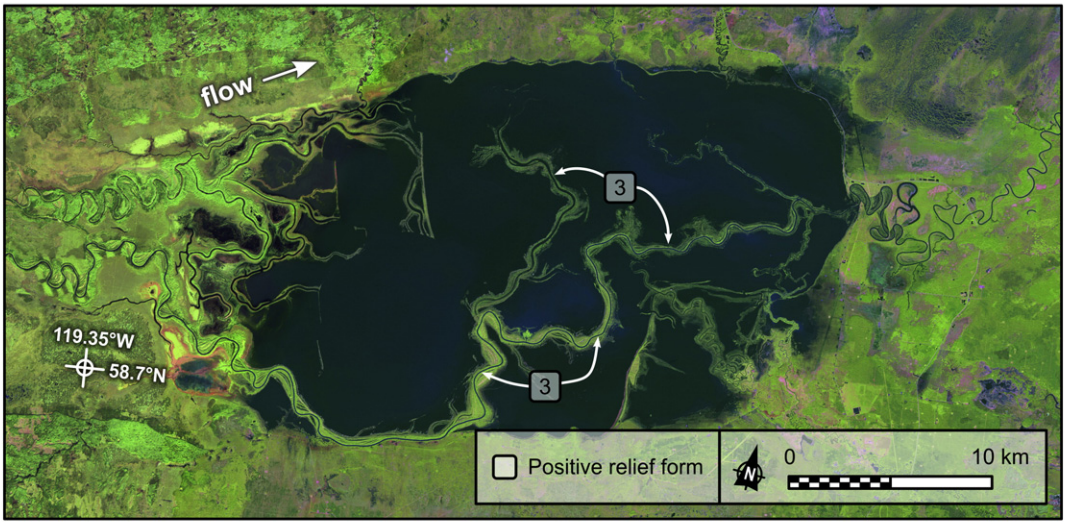


**Fig. 15.** The Hay River and Lake Zama, Alberta (58° 48′ N, 119° 06′ W) showing 'snail-trail' of prograding channel across lake (including 'failed' or discontinued arms), with some lateral accretion in delta area. Image taken on 21 June 2008. Image courtesy of Government of Canada, Natural Resources Canada, Earth Sciences Sector, Mapping Information Branch, Centre for Topographic Information — Sherbrooke SPOT4.

Alexander, 1999). The rate of sedimentation as typically quantified from modern and ancient environments of different types varies by at least eleven orders of magnitude (Sadler, 1981) and decreases as a power-law function of the interval of time over which it is measured (Strauss and Sadler, 1989; Tipper, 1983). The filling of negative relief by both autogenic and allocyclic-driven processes is episodic with non-depositional gaps of inactivity (hiatuses) and/or bursts of sedimentation (Ager, 1973; Macklin et al., 2010). Within individual elements of filled negative relief, the preserved stratigraphic record may be contaminated or overprinted by internally-generated and variable autogenic processes (Jerolmack and Paola, 2007; Straub et al., 2009; Van De Wiel and Coulthard, 2010; Wang et al., 2011a).

Direct fluvial input in rheic zones is through both bedload and finer sediment; in perirheic zones, infilling is largely by finer materials. The physical characteristics of sediment loads vary considerably. Up to 60% of the suspended sediment load of the middle Hwang He in China may be composed of sand-sized particles (Walling and Moorhead, 1989), thus supporting near-channel deposition and the creation of levees and alluvial ridges. High-latitude rivers may have low suspended loads resulting from lower runoff, lesser quantities of (fine) eroded soils from wetland-blanketed domains, and a proportionately greater channel reworking of (coarser) glacigenic and frost weathered material. Active lateral channel mobility of both meandering and braiding streams during peak melt discharges may create voids which are then only very slowly filled. In forest biomes floodplain depressions may also have a very long life where deforestation has not produced large quantities of fine sediment (Caselius, 1971). By contrast, in settled mid-latitudes soil erosion has contributed to blanket-like and accelerated fine overbank deposition, as for example in peak periods of agricultural activity in mediaeval Europe (Hoffmann et al., 2007; Macklin et al., 2010). Where such sediment has moved to historically hasdepended on delivery pathways and storages (Lang et al., 2003; James, 2013), with a proportion having yet to reach floodplains. Post-Pleistocene coarse sediment transport and channel mobility in temperate lowland rivers have been relatively restricted by interglacially-reduced stream power. In recent decades and centuries, complex response histories of river regulation have also involved channel engineering, floodplain inundation control, the trapping of fine sediment in reservoirs, and restricted channel dynamics because of a reduction in flows competent to transport bed material (Hudson et al., 2008; Lewin, 2013).

This is true on a global scale and for large floodplains where it affects the sedimentation potential of negative relief elements. Over millennial timescales accelerated soil erosion has increased inputs of fine sediment (Syvitski et al., 2005; Kundzewicz et al., 2009). Finer sediment from top-soil removal, towards the colloidal end of the river suspended sediment load, is more readily dispersed over floodplains far distant from channel margins unless moving in aggregate form (Walling and Moorhead, 1989). Secondly, and over the last 50 years in particular, large dam construction has reduced mainstream peak flows and transport-effective discharges, and trapped inputs of fine sediment from up-catchment (Nilsson et al., 2005; Gupta et al., 2012). Although there are wide differences between quantitative estimates of the amounts of accelerated erosion and dam storage, and therefore effects on downstream floodplains, centuries of accelerated sediment supply in some systems appear likely to be succeeded by ongoing decades of decrease (Syvitski et al., 2005; Walling, 2008; Wang et al., 2011b).

From a geomorphological perspective, and for quasi-natural floodplains, it is possible to begin constructing a qualitative framework for negative-relief 'survival-likelihood' (Table 3). In each of the domains initially recognised above (Table 1(i–iii)) negative relief is liable to differential elimination. To these should be added: (iv) under high river mobility, there may be reworking and elimination of whole floodplain assemblages, positive and negative elements alike; and (v) there may be relief preservation because of whole-assemblage isolation from mainstream reworking and infilling. That arises through river incision and terracing, or channel metamorphosis to a less active and laterally extensive state (Lewin and Macklin, 2003). Evidence for 'natural' survival rates is briefly discussed for each of these five groups in turn, although as yet global and corridor-variable survival rates have attracted limited study despite their biological significance.

**Table 3**
Natural negative-relief survival likelihood on large river floodplains (see text for explanation).

| | Negative relief form | Survival likelihood |
|---|---|---|
| (i) | Active channel segments | Dependent on channel branch dynamics |
| (ii) | Transitional zones with surface-water channel connection | Low |
| (iii) | Perirheic depressions | High |
| (iv) | Autogenic assemblage recycling | Restricted on large rivers |
| (v) | Metamorphosis/incision preservation | Possible in the long-term |

### 8.1. Segments of the channel network

Research has demonstrated that in both braided and anastomosing channel systems not all water-transmitting channels are geomorphologically active to the same degree (Makaske et al., 2009; Ashmore et al., 2011). For example, typically about 40% of experimental braid channels have active sediment transport at any one time (Bertoldi et al., 2009; Egozi and Ashmore, 2009). The same principle applies to plural large-river systems with mixed channel-patterning styles (Lewin and Ashworth, 2013). Sediment-transporting and eroding branches may be turbid and bed-abrasive, whereas geomorphologically moribund channels provide more sheltered habitats that are more suited as feeding and migration passages. How long moribund branches remain open has attracted limited documentation; most analyses concentrate on the causes and stability of bifurcation/avulsion (Huang and Nanson, 2007; Kleinhans et al., 2011). The lifespan of bifurcations varies greatly — from channel abandonment caused by switching or plugging in a single flood event, to the 100s to 1000s of years it can take for a river to change its course on a floodplain or delta (Kleinhans et al., 2013).

### 8.2. Transitional zones still with surface-water channel connection

Many floodplain channels in the transitional zone of large rivers are geomorphologically moribund but hydrologically conjoined to the main trunk stream. Channels can remain open so long as they can be scoured by clear water returning to the trunk stream during lower flows. If the supply of fine sediment is adequate, advective sediment transport is driven into connected backwater zones and high rates of deposition are likely there (Day et al., 2008; Citterio and Piégay, 2009). Because so many large river systems anabranch dynamically in plural styles (Latrubesse, 2008; Lewin and Ashworth, 2013), it follows that a range of partially connected or disconnected Y-junction channel forms should be expected, extending further away from trunk streams (Trigg et al., 2012; Kleinhans et al., 2013). Infilling involves a whole range of negative relief elements of different origin (Dunne and Aalto, 2013; Lewin and Ashworth, 2013) including mainstem bifurcations, formerly active channel segments still mainstem-connected at one end, branching-off accessory channels, downstream return channel junctions, and tributary input channels. These may undergo active erosion/sedimentation, or they may transmit passively, back-up, or pond water prior to slow sediment infilling. As yet there is little quantitative data for sedimentation rates for the filling of partially disconnected channel segments along large rivers. Latrubesse and Franzinelli (2002) note an abandoned channel belt on the Amazon that is still incompletely filled after c.1000 years. Trigg et al. (2012) conducted bed sonar surveys on 33 floodplain channels connected to the Amazon trunk stream that ranged from 40 to 900 m width. Their work showed a consistent relationship between depth of the floodplain channels and the water source. Channels with no local runoff input had a mean depth of 10.4 m, those with runoff input from the local catchment had a mean depth of 15.9 m and those channels strongly connected to the main river had a mean depth of 17.7 m. Clearly, despite the high sediment loads associated with many large rivers, infill rates are slow, at least in partially or seasonally, hydrologically-connected floodplain channels.

### 8.3. Perirheic forms

By contrast, negative relief elements (and many lentic environments) that are distant from, and poorly linked to, main channels may have extended lives with very low sedimentation rates (Citterio and Piégay, 2009). This greatly depends on inundation patterns and depths, and upon floodwater sediment loadings. On some large rivers, such as the Strickland (Aalto et al., 2008), levee growth of up to 2 m can confine high flows to the main channel so that accretion on the floodplain is shut down for decades until a sufficiently large flood overtops the levee (Dunne and Aalto, 2013).

Field surveys have focused on 'oxbow' cutoff infilling (Werritty et al., 2006; Citterio and Piégay, 2009; Toonen et al., 2012) and, more generally, on dispersed rates of overbank sedimentation (Asselman and Middelkoop, 1995; Walling et al., 1996; Törnqvist and Bridge, 2002; Benedetti, 2003; Aalto and Nittrouer, 2012). Several overbank sedimentation rate models have been proposed (reviewed in Marriott, 1996) with deposition rates and the size of deposited material decreasing with distance from channels. In practice measured deposition rates are locally complex, with high rates near channel margins and in relation to floodplain microtopography. Domination of near-channel fine sediment accretion is consistent with those surveys that show that marginal and 'inner accretionary bank' accumulation on point bars rapidly follows channel migration (Steiger et al., 2001; Gautier et al., 2010). Averaged over time, deposition rates (of sand in particular) may decrease abruptly away from channels, although this can lead to more rapid channelway aggradation and an enhanced likelihood for avulsion (Törnqvist and Bridge, 2002). Avulsive splays, dryland floodouts, depositional lobes or inland deltas on a range of scales may partially infill occluded basins (Latrubesse and Franzinelli, 2002; Ramonell et al., 2002: Assine and Silva, 2009; Latrubesse et al., 2010, their Fig. 8; Tooth, 2013). Along the Amazon and the middle Paraná in particular, lakes behind channel belt barriers are only partially filled by sediment splays even when close to main channels.

### 8.4. Autogenic assemblage recycling

Only a few large trunk rivers like the Jamuna/Brahmaputra channel belt (Ashworth and Lewin, 2012) or the Indus (Fig. 11) recycle their entire floodplains over decades to centuries. Some, like reaches of the Paraná and Amur Rivers, have braided main channels set within wetlands and actively meandering accessory channel systems. The Lower Mississippi has actively reworked its floodplain so that few meander trains persist for longer than 120 years (Harmar and Clifford, 2006) although the river has been subject to multiple Holocene channel-belt avulsions. The Ganga exhibits decade-scale meander cycles (Gupta et al., 2013) within a longer-term history of valley-floor response to monsoonal oscillation (Roy et al., 2011). Other rivers like the Amazon (Mertes et al., 1996; Rozo et al., 2012) have quasi-straight reaches that, without bend development, oscillate laterally over time to produce bordering multi-ridge accretionary zones of limited width. More laterally-remote wetlands and floodplains on large rivers frequently do not appear to be effectively reworked on a centuries to millennial timescale.

### 8.5. Metamorphosis or incisional preservation

Many rivers have terrace flights that preserve palaeoforms on their surfaces. On larger cratonic rivers, Quaternary depths of valley incision may be slight (Bridgland and Westaway, 2008), to the extent that palaeoforms relating to prior climatic regimes remain at near-channel level. With restricted lateral geomorphological reworking, elements that were geomorphologically active several thousand years previously may still form part of the inundated floodplain (Bravard, 1989; Sidorchuk, 2003; Busschers et al., 2007; Erkens et al., 2009; Latrubesse, 2010; Lewin and Ashworth, 2013; Valente et al., 2013). Likewise, palaeovalley systems may preserve a significant morphological expression that is buried beneath Holocene delta plains (Busschers et al., 2007; Fontana et al., 2008; Blum et al., 2013).

## 9. Management considerations

In the last several centuries, geomorphological processing in major floodplains has been transformed through human activity, for example through flow regulation, channelization, land drainage and for flood

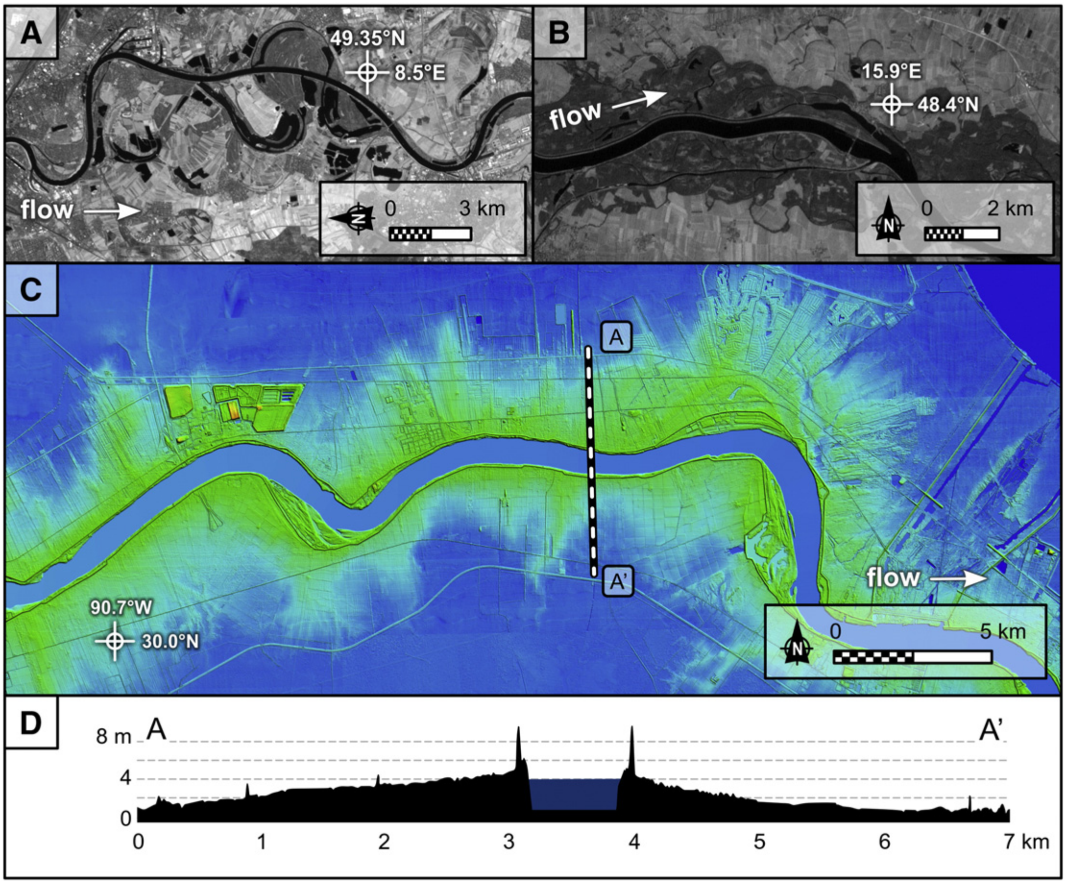


**Fig. 16.** (A) The channelized and formerly meandering Rhine, Germany (49° 25′ N, 8° 30′ E). Image taken on 15 April 2003; (B) the regulated, channelized and formerly anabranching Danube, Austria (48° 23′ N, 15° 51′ E). Image taken on 27 March 2003; (C) LiDAR image of the Mississippi between New Orleans and Baton Rouge with both artificial and natural flood embankments (D) showing detachment of river from its floodplain. Image from 2001 and courtesy of Digital Elevation Model (USGS DEM), SW quadrant of Reserve quadrangle, Louisiana, UTM 15 NAD83, Louisiana FEMA Project — Phase 1: Lake Maurepas/Lake. All Landsat imagery courtesy of the U.S. Geological Survey.

protection (Hudson et al., 2008) (Fig. 16). The scale and nature of sediment recycling has been considerably modified to homogenize ecosystems (van de Wolfshaar et al., 2011) and to fossilize alluvial dynamics (Hohensinner et al., 2011). Under natural conditions, relief forms are diachronous and have different rates of element (re)formation. If particular negative forms are now to be preserved or restored/repaired in order to maintain hydromorphological quality, it must be appreciated that their 'normal' development is no longer in place. Like biological entities they are dynamic, albeit over varying and generally much longer timescales. Whilst allowing for short-term 'adjustments' (as evidenced by eroding banks or silting beds), many hydromorphic approaches to river systems appear to regard the morphological 'platform' as immutable until modified by human agency. This is not the reality that an analysis of negative relief forms reveals.

Three examples may be used to illustrate this. First, slackwater arms in rheic transition zones bordering contemporary channels are part of an evolving channel network. Their biological significance – for example as refuges, or for feeding or spawning – may lead to their artificial preservation or restoration. Ecologically useful though this may be, the long-term existence of such features is as components of dynamic systems that produce sequences of active and relict channels, which are then filled with sediment. Such arms may not individually have a long-term natural existence. Second, cutoff 'oxbow' lakes are produced through channel migration and then slowly infilled subsequently. These may have an extended 'life', but static preservation only of the floodplain set that currently exists is a distortion of their natural replacement sequence. Third, major wetlands may have existed for at least $10^4$ years repeatedly during the Quaternary as occluding sediment bodies have built up or subsidence has occurred; once drained, artificial restoration would require both major engineering and attention to the intricacies of their former water drainage and sedimentation nets that evolved over millennia.

It may reasonably be argued that recognition of the ecological significance of the totality of negative relief forms, and preservation/restoration of some of them, are very worthwhile objectives (Hohensinner et al., 2011). But it has equally to be appreciated that in practice static conservation of form elements can only achieve 'half-natural' states. Opening-up water-filled voids and 'allowing the river to do the work' may underestimate both the timescale which such work will require and the varied cycles of formation and infilling that produced observed sets of features. The fidelity with which forms, conceived of as being static, can be artificially maintained is limited.

## 10. Conclusions

Negative relief on larger floodplains – important to hydrology, biology and geology, and thus to humans – is highly varied in terms of dimensions and dynamics. This paper shows that modification of formerly active channels and the preservation of negative relief elements on large floodplains is more complex than can be captured by a twofold channel/overbank descriptive framework. In *perirheic zones*, there are palaeoforms inherited from prior morphogenetic systems and ones related to tectonic depressions. Though geomorphologically relatively permanent and inactive, these may still be connected to active hydrobiological systems. By contrast, hydrologically-linked *transitional zones* at rheic margins are the scene for much varied geomorphologically activity. This zone performs especially significant transformational roles in modifying *rheic zone* channel forms, but ones that are more complex than the now relatively understood channel bar, cutoff and avulsion processes.

In specific terms, this paper has suggested that patterns of channel-margin sedimentation in partially-abandoned or flow-detached channel segments, and with low or discontinuous mainstream channel banks, are important. Geomorphologically-active *rheic zones* may occupy only part of the water system that provides hydrobiological diversity; branches and backwaters may or may not be historical hangovers now being modified by ongoing geomorphological processes. Prior surface water ponding across floodplains may aid negative relief preservation by inhibiting the lateral dispersal and sequestration of sediment, whilst vegetation may focus sediment sequestration. Braid and meander floodplains differ because of the ways their channels are abandoned and incorporated into floodplains as well as because of contrasts between the initial rheic patterns themselves. Sinuous channels prograding into extensive wetlands and shallow waters may form rather differently from those created by lateral meander mobility.

Still- and slack-water depressions and channel arms are both biologically significant yet also particularly sensitive to anthropogenic activities. These have included millennial-scale land cover changes, century-scale channel impoundment and engineering, and contemporary induced climatic change. At and beyond channel margins and away from mainstem flows, it is relatively easy to adopt sustainable engineering solutions for both hazard protection and economic development. But this may produce a reduction in geodiversity, which inevitably also leads on to a reduction of biodiversity. Although slow to operate (and over mixed timescales), the geomorphological processes and replacement cycles that generate floodplain negative relief variety continue to require further research. Future research needs to enhance: (i) the quantitative archive of large floodplain element hierarchies for a greater variety of large floodplains, especially using data derived from satellite imagery; (ii) using both field survey, historical archives, (maps and images) and geochronology, more comprehensive knowledge of the dynamics and survival spans under both quasi-natural and anthropogenically-modified conditions.

The management of negative relief on large river floodplains needs to quantify *both* the geomorphological and hydrological connectivity between river and floodplain. Understanding both ecosystems and the distribution of organic resources requires knowledge of the alluvial process context for negative relief formation and preservation.

## Acknowledgements

PJA would like to thank the UK Natural Environment Research Council (NERC) for support under grant NE/E016022/1 (awarded to Ashworth, Hardy, Lane, Nicholas, Parsons and Sambrook Smith with Best) that funded fieldwork on the Rio Paraná, Argentina and was the catalyst for stimulating discussions between JL and PJA on the controls on large river dynamics. We are extremely grateful to Chris Simpson of Fulcrum Graphic Communications Inc., Vancouver, B.C., for his expertise in sourcing and producing the figures. Derald Smith kindly allowed us to use his image in Fig. 1C. Fig. 12 was produced from aerial sorties commissioned by PJA, Jim Best and Greg Sambrook Smith. We are especially grateful to Kim Cohen, Edgardo Latrubesse, Andrew Nicholas and an anonymous referee for their considerable help in improving the text.

## References

Aalto, R., Nittrouer, C.A., 2012. $^{210}$Pb geochronology of flood events in large tropical river systems. Phil. Trans. R. Soc. A 370, 2040–2074. http://dx.doi.org/10.1098/rsta.2011.0607.

Aalto, R., Maurice-Bourgoin, L., Dunne, T., Montgomery, D.R., Nittrouer, C.A., Guyot, J.L., 2003. Episodic sediment accumulation on Amazonian floodplains influenced by El Niño/Southern Oscillation. Nature 425, 493–497. http://dx.doi.org/10.1038/nature02002.

Aalto, R., Lauer, J.W., Dietrich, W.E., 2008. Spatial and temporal dynamics of sediment accumulation and exchange along Strickland River floodplains (Papua New Guinea) over decadal-to-centennial timescales. J. Geophys. Res. 113, F01S04. http://dx.doi.org/10.1029/2006JF000627.

Adams, P.N., Slingerland, R.L., Smith, N.D., 2004. Variations in natural levee morphology in anastomosed channel flood plain complexes. Geomorphology 61, 127–142.

Ager, D.V., 1973. The Nature of the Stratigraphic Record. Wiley, New York (114 pp.).

Alho, C.J.R., Lacher, T.E., Gonçalves, H.C., 1988. Environmental degradation in the Pantanal ecosystem. Bioscience 38, 164–171.

Archer, D.R., 2005. Review of Amazonian depositional systems. In: Blum, M.D., Marriott, S.B., Leclair, S.F. (Eds.), Fluvial Sedimentology VII. Special Publication 35, International Association of Sedimentologists. Blackwell, Oxford, pp. 17–39.

Ashmore, P.E., 1991. How do gravel bed streams braid? Can. J. Earth Sci. 28, 326–341.

Ashmore, P., Bertoldi, W., Gardner, J.T., 2011. Active width of gravel-bed braided rivers. Earth Surf. Process. Landf. 36, 1510–1521. http://dx.doi.org/10.1002/esp.2182.

Ashworth, P.J., Lewin, J., 2012. How do big rivers come to be different? Earth Sci. Rev. 114, 84–107. http://dx.doi.org/10.1016/j.earscirev.2012.05.003.

Ashworth, P.J., Best, J.L., Peakall, J., Lorsong, J.A., 1999. The influence of aggradation rate on braided alluvial architecture: field study and physical scale modelling of the Ashburton River gravels, Canterbury Plains, New Zealand. In: Smith, N.D., Rogers, J. (Eds.), Fluvial Sedimentology VI, Special Publication. International Association of Sedimentologists, 28. Blackwell, Oxford, pp. 333–346.

Ashworth, P.J., Best, J.L., Roden, J.E., Bristow, C.S., Klaassen, G.J., 2000. Morphological evolution and dynamics of a large, sand braid-bar, Jamuna River, Bangladesh. Sedimentology 47, 533–555.

Ashworth, P.J., Sambrook Smith, G.H., Best, J.L., Bridge, J.S., Lane, S.N., Lunt, I.A., Reesink, A.J.H., Simpson, C.J., Thomas, R.E., 2011. Evolution and sedimentology of a channel fill in the sandy braided South Saskatchewan River and its comparison to the deposits of an adjacent compound bar. Sedimentology 58, 1860–1883. http://dx.doi.org/10.1111/j.1365-3091.2011.01242.x.

Asselman, N.E.M., Middelkoop, H., 1995. Floodplain sedimentation: quantities, patterns and processes. Earth Surf. Process. Landf. 20, 481–499.

Assine, M.L., Silva, A., 2009. Contrasting fluvial styles of the Paraguay River in the northwestern border of the Pantanel wetland, Brazil. Geomorphology 113, 189–199. http://dx.doi.org/10.1016/j.geomorph.2009.03.012.

Assine, M.L., Soares, P., 2004. Quaternary of the Pantanal, west-central Brazil. Quat. Int. 114 (1), 23–34.

Aufdenkamp, A.L., Mayorga, E., Raymond, P.A., Melack, J.M., Doney, S.C., Alin, S.R., Aalto, R.E., Yoo, K., 2011. Riverine coupling of biogeochemical cycles between land, oceans and atmosphere. Ecol. Environ. 9, 53–60. http://dx.doi.org/10.1890/100014.

Bastviken, D., Tranvik, L.J., Downing, J.A., Crill, P.M., Enrich-Prast, A., 2011. Freshwater methane emissions offset the continental carbon sink. Science 331, 50. http://dx.doi.org/10.1126/science.1196808.

Benedetti, M.M., 2003. Controls on overbank deposition in the upper Mississippi River. Geomorphology 56, 271–290.

Bertoldi, W., Ashmore, P., Tubino, M., 2009. A method for estimating the mean bed load flux in braided rivers. Geomorphology 103, 330–340. http://dx.doi.org/10.1016/j.geomorph.2008.06.014.

Best, J.L., Ashworth, P.J., 1997. Scour in large braided rivers and the recognition of sequence stratigraphic boundaries. Nature 387, 275–277.

Best, J.L., Ashworth, P.J., Bristow, C.S., Roden, J.E., 2003. Three-dimensional sedimentary architecture of a large, mid-channel sand braid bar, Jamuna River, Bangladesh. Journal of Sedimentary Research 73, 516–530.

Best, J., Woodward, J., Ashworth, P., Sambrook Smith, G., Simpson, C., 2006. Bar-top hollows: A new element in the architecture of sandy braided rivers. Sedimentary Geology 190, 241–255.

Blondeaux, P., Seminara, G., 1985. A unified bar-bend theory of river meanders. J. Fluid Mech. 112, 363–377.

Blum, M., Martin, J., Milliken, K., Garvin, M., 2013. Paleovalley systems: Insights from Quaternary analogs and experiments. Earth Sci. Rev. 116, 128–169. http://dx.doi.org/10.1016/j.earscirev.2012.09.003.

Bravard, J.-P., 1989. La métamorphose des rivières des Alpes français à la fin du Moyen-Age et à l'époque moderne. Bull. Soc. Géog. Liège 25, 145–157.

Bridge, J.S., 2003. Rivers and Floodplains: Forms, Processes and Sedimentary Record. Blackwell, Malden, USA (491 pp.).

Bridge, J.S., Lunt, I.A., 2006. Depositional models of braided rivers. In: Sambrook Smith, G.H., Best, J.L., Bristow, C.S., Petts, G.E. (Eds.), Braided Rivers; Process, Deposits, Ecology and Management, 36. International Association of Sedimentologists Special Publication, Blackwell, Oxford, pp. 11–50.

Bridge, J.S., Alexander, J., Collier, R.E.LL, Gawthorpe, R.L., Jarvis, J., 1995. Ground-penetrating radar and coring used to study the large-scale structure of point-bar deposits in 3-D. Sedimentology 42 (6), 839–852.

Bridgland, D.R., Westaway, R., 2008. Climatically controlled river terrace staircases: a worldwide Quaternary phenomenon. Geomorphology 98, 288–315.

Brierley, G., 1991. Bar sedimentology of the Squamish River, British Colombia: definition and application of morphostratigraphic units. J. Sediment. Res. 61, 211–225. http://dx.doi.org/10.1306/D42676D6-2B26-11D7-8648000102C1865D.

Brooks, G.R., 2003. Alluvial deposits of a mud-dominated stream: the Red River, Manitoba, Canada. Sedimentology 50, 441–458.

Brown, A.G., Toms, P.S., Carey, C.J., Howard, A.J., Challis, K., 2013. Late Pleistocene–Holocene river dynamics at the Trent-Soar confluence, England, UK. Earth Surf. Process. Landf. 38, 237–249. http://dx.doi.org/10.1002/esp.3270.

Bullard, J.E., McTainsh, G.H., 2003. Aeolian-fluvial interactions in dryland environments: examples, concepts and Australia case study. Prog. Phys. Geogr. 27, 471–501.

Busschers, F.S., Kasse, C., van Balen, R.T., Vandenberghe, J., Cohen, K.M., Weerts, H.J.T., Wallinga, J., Johns, C., Cleveringa, P., Bunnik, F.P.M., 2007. Late Pleistocene evolution of the Rhine–Meuse system in the southern North Sea basin: imprints of climate change, sea level oscillation and glacio-isostacy. Quat. Sci. Rev. 26, 3216–3248.
Caselius, C., 1971. On the meandering rivers in the Lokka reservoir district, Finland. Fennia 90, 1–71.
Castro, N.M., Hornberger, G.M., 1991. Surface-subsurface water interactions in an alluviated mountain stream channel. Water Resour. Res. 27, 1613–1621.
Cheetham, G.H., 1979. Flow competence in relation to stream channel form and braiding. Geol. Soc. Am. Bull. 90, 877–886.
Citterio, A., Piégay, H., 2009. Overbank sedimentation rates in former channel lakes: characterization and control factors. Sedimentology 56, 461–482. http://dx.doi.org/10.1111/j.1365-3091.2008.00979.x.
Constantine, J.A., Dunne, T., 2009. Meander cutoff and the controls on the production of oxbow lakes. Geology 36, 23–26. http://dx.doi.org/10.1130/G24130A.1.
Constantine, J.A., Dunne, T., Piégay, H., Kondolf, G.M., 2010. Controls on the alluviation of oxbow lakes by bed-material load along the Sacramento River, California. Sedimentology 57, 389–407. http://dx.doi.org/10.1111/j.1365-3091.2009.01084.x.
Day, G., Dietrich, W.E., Rowland, J.C., Marshall, A., 2008. The depositional web on the floodplain of the Fly River, Papua New Guinea. J. Geophys. Res. 113, F01S02. http://dx.doi.org/10.1029/2006JF000622.
Dieras, P.L., Constantine, J.A., Hales, T.C., Piégay, H., Riquier, J., 2013. The role of oxbow lakes in the off-channel storage of bed material along the Ain River, France. Geomorphology. http://dx.doi.org/10.1016/j.geomorph.2012.12.024.
Dietrich, W.E., Day, G., Parker, G., 1999. The Fly River, Papua New Guinea: inferences about river dynamics, floodplain sedimentation and fate of sediment. In: Miller, A., Gupta, A. (Eds.), Varieties of Fluvial Form. Wiley, Chichester, pp. 345–376.
Doeschl, A.B., Ashmore, P., Davison, M., 2006. Methods for assessing exploratory computational models of braided rivers. In: Sambrook Smith, G.H., Best, J.L., Bristow, C.S., Petts, G.E. (Eds.), Braided Rivers: Process, Deposits, Ecology and Management. International Association of Sedimentologists Special Publication, 36. Blackwell, Oxford, pp. 177–197.
Dulal, K.P., Shimizu, Y., 2010. Experimental simulation of meandering in clay mixed sediments. J. Hydro Environ. Res. 4, 329–343. http://dx.doi.org/10.1016/j.jher.2010.05.001.
Dunne, T., Aalto, R.E., 2013. Large river floodplains. In: Shroder, J.F. (Ed.), Treatise on Geomorphology, vol. 9. Academic Press, San Diego, CA, pp. 645–678. http://dx.doi.org/10.1016/B978-0-12-374739-6.00258-X.
Dunne, T., Mertes, L.A.K., Meade, R.H., Richey, J.E., Forsberg, B.R., 1998. Exchanges of sediment between the flood plain and the channel of the Amazon River in Brazil. Geol. Soc. Am. Bull. 110, 450–467.
Egozi, R., Ashmore, P., 2009. Experimental analysis of braided channel pattern response to increased discharge. J. Geophys. Res. Earth Surf. 114, F02012. http://dx.doi.org/10.1029/2008JF001099.
Erkens, G., Dambeck, R., Vollenberg, K.P., Bouman, M.T.I.J., Boss, J.A.A., Cohen, K.M., Wallinga, J., Hoek, W.Z., 2009. Fluvial terrace formation in the northern Upper Rhine Graben during the last 20 000 years as a result of allogenic controls and autogenic evolution. Geomorphology 103, 476–495.
Erkens, G., Hoffmann, T., Gerlach, R., Klostermann, J., 2011. Complex fluvial response to Lateglacial and Holocene allogenic forcing in the Lower Rhine Valley (Germany). Quat. Sci. Rev. 30, 611–627. http://dx.doi.org/10.1016/j.quascirev.2010.11.019.
Ferguson, R.J., Brierley, G.J., 1999. Levee morphology and sedimentology along the lower Turos River, south-eastern Australia. Sedimentology 46, 627–648.
Fontana, A., Mozzi, P., Bondesan, A., 2008. Alluvial megafans in the Venetian–Friulian Plain (north-east Italy): evidence of sedimentary and erosive phases during Late Pleistocene and Holocene. Quat. Int. 189, 71–90.
Fraser, L.H., Keddy, P.A., 2005. The World's Largest Wetlands: Ecology and Conservation. Cambridge University Press, Cambridge.
Fryirs, K., Brierley, G.J., 2001. Variability in sediment delivery and storage along river courses in Bega catchment, NSW, Australia: implications for re-mobilization of stored in channel sediment. Geomorphology 38, 237–265.
Fustic, M., Hubbard, S.M., Spencer, R., Smith, D.G., Leckie, D.A., Bennett, B., Larter, S., 2012. Recognition of down-valley translation in tidally influenced meandering fluvial deposits, Athabasca oil sands (Cretaceous), Alberta, Canada. Mar. Pet. Geol. 29, 219–232. http://dx.doi.org/10.1016/j.marpetgeo.2011.08.004.
Gardner, J.T., Ashmore, P.E., 2011. Geometry and grain-size characteristics of the basal surface of a braided river deposit. Geology 39 (3), 247–250. http://dx.doi.org/10.1130/G31639.1.
Gautier, E., Brunstein, D., Vauchel, P., Jouanneau, J.-M., Roulet, M., Garcia, C., Guyot, J.-L., Castro, M., 2010. Channel and floodplain sediment dynamics in a reach of the tropical meandering Rio Beni (Bolivian Amazonia). Earth Surf. Process. Landf. 35, 1838–1853. http://dx.doi.org/10.1002/esp.2065.
Gay, G.R., Gay, H.H., Gay, W.H., Martinson, H.A., Meade, R.H., Moody, J.A., 1998. Evolution of cutoffs across meander necks in Powder River, Montana, USA. Earth Surf. Process. Landf. 19, 651–662.
Gibbard, P.L., Lewin, J., 2002. Climate and related controls on interglacial fluvial sedimentation in lowland Britain. Sediment. Geol. 151, 187–210.
Grenfell, M., Aalto, R., Nicholas, A., 2012. Chute channel dynamics in large, sand-bed meandering rivers. Earth Surf. Process. Landf. 37, 315–331. http://dx.doi.org/10.1002/esp.2257.
Gupta, A. (Ed.), 2007. Large Rivers: Geomorphology Management. Wiley, Chichester.
Gupta, H., Kao, S.-J., Dai, M., 2012. The role of mega dams in reducing sediment fluxes: a case study of large Asian rivers. J. Hydrol. 464–5, 447–458. http://dx.doi.org/10.1016/j.hydrol.2012.07.038.
Gupta, N., Atkinson, P.M., Carling, P.A., 2013. Decadal length changes in the fluvial planform of the River ganga: bringing a mega-river to life with Landsat archives. Remote Sens. Lett. 4, 1–9. http://dx.doi.org/10.1080/2150704X.2012.682658.
Gurnell, A.M., Surian, N., Zanoni, L., 2009. Multi-thread river channels: a perspective on changing European alpine river systems. Aquat. Sci. 71, 253–265.
Gurnell, A.M., Bertoldi, W., Corenblit, D., 2012. Changing river channels: the roles of hydrological processes, plants and pioneer fluvial landforms. Earth Sci. Rev. 111 (1–2), 129–141. http://dx.doi.org/10.1016/j.earscirev.2011.11.005.
Ham, D., Church, M., 2012. Morphodynamics of an extended bar complex, Fraser River, British Columbia. Earth Surf. Process. Landf. 37, 1074–1089. http://dx.doi.org/10.1002/esp.3231.
Harmar, O.P., Clifford, N.J., 2006. Planform dynamics of the Lower Mississippi River. Earth Surf. Process. Landf. 31, 825–843.
Hijma, M.P., Cohen, K.M., 2011. Holocene transgression of the Rhine river mouth area, The Netherlands/Southern North Sea: palaeogeography and sequence stratigraphy. Sedimentology 58, 1453–1485. http://dx.doi.org/10.1111/j.1365-3091.2010.01222.x.
Hoffmann, T., Erkins, G., Cohen, K.M., Houben, P., Seidel, J., Dikau, R., 2007. Holocene floodplain sediment storage and hillslope erosion within the Rhine catchment. The Holocene 17, 105–118.
Hoffmann, T., Erkins, G., Gerlach, R., Klostermann, J., Lang, A., 2009. Trends and controls of Holocene floodplain sedimentation in the Rhine catchment. Catena 77, 96–106.
Hohensinner, S., Jungwirth, M., Muhar, S., Schmutz, S., 2011. Spatio-temporal habitat dynamics in a changing Danube River landscape 1812–2006. River Res. Appl. 27, 939–955. http://dx.doi.org/10.1002/rra.1407.
Hooke, J.M., Yorke, L., 2011. Channel bar dynamics on multi-decadal timescales in an actively meandering river. Earth Surf. Process. Landf. 36, 1910–1928. http://dx.doi.org/10.1002/esp.2214.
Horn, J.D., Fielding, C.R., Joeckel, R.M., 2012. Revision of Platte River alluvial facies model through observations of extant channels and barforms, and subsurface alluvial valley fills. J. Sediment. Res. 82, 72–91.
Howard, A.D., 1984. Simulation model of meandering. In: Elliott, C.M. (Ed.), River Meandering: Proceedings of the Conference Rivers' 83. American Society of Civil Engineers, New York, pp. 952–963.
Huang, H.Q., Nanson, G.C., 2007. Why some alluvial rivers develop an anabranching pattern. Water Resour. Res. 43, W07441. http://dx.doi.org/10.1029/2006WR005223.
Hubbard, S.M., Smith, D.G., Nielsen, H., Leckie, D.A., Fustic, M., Spencer, R.J., Bloom, L., 2011. Seismic geomorphology and sedimentology of a tidally influenced river deposit, Lower Cretaceous Athabasca oil sands, Alberta, Canada. Am. Assoc. Pet. Geol. 95, 1123–1145. http://dx.doi.org/10.1306/12131010111.
Hudson, P.F., Middlekoop, H., Stouthamer, E., 2008. Flood management along the Lower Mississippi and Rhine Rivers (The Netherlands) and the continuum of geomorphic adjustment. Geomorphology 101, 209–236. http://dx.doi.org/10.1016/j.geomorph.2008.07.001.
Iriondo, M., 2004. Large wetlands of South America: a model for Quaternary humid environments. Quat. Int. 114, 3–9. http://dx.doi.org/10.1016/S1040-6182(03)00037-5.
Iriondo, M.H., Paggi, J.C., Parma, M.J. (Eds.), 2007. The Middle Paraná River. Springer, Verlag: Berlin.
James, L.A., 2013. Legacy sediment: definitions and processes of episodically produced anthropogenic sediment. Anthropocene. http://dx.doi.org/10.1016/j.ancene.2013.04.001.
Jerolmack, D., Paola, C., 2007. Complexity in a cellular model of river avulsion. Geomorphology 91, 259–270. http://dx.doi.org/10.1016/j.geomorph.2007.04.022.
Kemp, J., Spooner, N.A., 2007. Evidence for regionally wet conditions before the LGM in southeast Australia: OLS ages from a large palaeochannel in the Lachlan Valley. J. Quat. Sci. 22, 423–427. http://dx.doi.org/10.1002/jqs.1125.
Khan, N.S., Horton, B.P., McKee, K.L., Jerolmack, D., Falcini, F., Enache, M.D., Vane, C.H., 2013. Tracking sedimentation from the historic A.D. 2011 Mississippi River flood in the deltaic wetlands of Louisiana, USA. Geology. http://dx.doi.org/10.1130/G33805.1.
Kleinhans, M.G., Cohen, K.M., Hoekstra, J., Ljmker, J.M., 2011. Evolution of a bifurcation in a meandering river with adjustable channel widths, Rhine delta apex, The Netherlands. Earth Surf. Process. Landf. 36, 2011–2027. http://dx.doi.org/10.1002/esp.2222.
Kleinhans, M.G., Ferguson, R.I., Lane, S.N., Hardy, R., 2013. Splitting rivers at their seams: bifurcations and avulsions. Earth Surf. Proc. Land. 38, 47–61. http://dx.doi.org/10.1002/esp.3268.
Kundzewicz, Z.W., Nohara, D., Tong, J., Oki, T., Buda, S., Takeuchi, K., 2009. Discharge of large Asian rivers — observations and projections. Quat. Int. 208, 4–10.
Lane, S.N., 2006. Approaching the system-scale understanding of braided river behaviour. In: Sambrook Smith, G.H., Best, J.L., Bristow, C.S., Petts, G.E. (Eds.), Braided Rivers: Process, Deposits, Ecology and Management. International Association of Sedimentologists Special Publication, 36. Blackwell, Oxford, pp. 107–135.
Lane, S.N., Westaway, R.M., Hicks, D.M., 2003. Estimation of erosion and deposition volumes in a large gravel-bed, braided river using synoptic remote sensing. Earth Surf. Process. Landf. 28, 249–271.
Lang, A., Bork, H.-R., Mäckel, R., Presto, N., Wunderlich, J., Dikau, R., 2003. Changes in sediment flux and storage within a fluvial system: some examples from the Rhine catchment. Hydrol. Process. 17, 3321–3334. http://dx.doi.org/10.1002/hyp.1389.
Latrubesse, E.M., 2008. Patterns of anabranching channels: the ultimate end-member adjustment of mega rivers. Geomorphology 101, 130–145. http://dx.doi.org/10.1016/j.geomorph.2008.05.035.
Latrubesse, E.M., 2010. Amazon lakes. In: Bengtsson, L., Herschy, R.W., Fairbridge, R.W. (Eds.), Encyclopedia of Lakes and Reservoirs. Springer, Netherlands, pp. 13–26.
Latrubesse, E.M., Franzinelli, E., 2002. Holocene alluvial plain of the middle Amazon River Brazil. Geomorphology 44, 241–257. http://dx.doi.org/10.1016/S0169-555X(01)00177-5.
Latrubesse, E.M., Franzinelli, E., 2005. The late Quaternary evolution of the Negro River, Amazon, Brazil: implications for island and floodplain formation in large anabranching tropical systems. Geomorphology 70, 372–397.
Latrubesse, E.M., Kalicki, T., 2002. Late Quaternary palaeohydrological changes in the upper Purus Basin, southwestern Amazonia, Brazil. Z. Geomorphol. Suppl. 129, 41–59.

Latrubesse, E.M., Stevaux, J.C., Sinha, R., 2005. Tropical rivers. Geomorphology 70, 187–206. http://dx.doi.org/10.1016/j.geomorph.2005.02.005.

Latrubesse, E.M., Cozzuel, M., da Silva-Caminha, S.A.F., Rigsby, C.A., Absy, M.L., Jaramillo, C., 2010. The Late Miocene palaeogeography of the Amazon Basin and the evolution of the Amazon River system. Earth Sci. Rev. 99, 99–124. http://dx.doi.org/10.1016/j.earscirev.2010.02.005.

Latrubesse, E.M., Stevaux, J.C., Cremon, E.H., May, J.-H., Tatumi, S.H., Hurtado, M.A., Bezada, M., Argollo, J.B., 2013. Late Quaternary megafans, fans and fluvio–aeolian interactions in the Bolivian Chaco, Tropical South America. Palaeogeogr. Palaeoclimatol. Palaeoecol. 356–357, 75–88. http://dx.doi.org/10.1016/j.palaeo.2012.04.003.

Lauer, J.W., Parker, G., 2008. Net local removal of floodplain sediment by river meander migration. Geomorphology 96, 123–149.

Leclerc, R.F., Hickin, E.J., 1997. The internal structure of scrolled floodplain deposits based on ground-penetrating radar, North Thompson River, British Columbia. Geomorphology 21, 17–38.

Leddy, J.O., Ashworth, P.J., Best, J.L., 1993. Mechanisms of anabranch avulsion within gravel-bed braided rivers: observations from a scaled physical model. In: Best, J.L., Bristow, C.S. (Eds.), Braided Rivers. Geological Society of London Special Publication, 75, pp. 119–127.

Lewin, J., 1983. Changes of channel pattern and floodplains. In: Gregory, K.J. (Ed.), Background to Palaeohydrology. Wiley, Chichester, pp. 303–319.

Lewin, J., 2010. Medieval environmental impacts and feedbacks: the lowland floodplains of England and Wales. Geoarchaeology 25, 267–311.

Lewin, J., 2013. Enlightenment and the GM floodplain. Earth Surf. Process. Landf. 38, 17–29. http://dx.doi.org/10.1002/esp.3230.

Lewin, J., Ashworth, P.J., 2013. Defining large river channel patterns: alluvial exchange and plurality. Geomorphology. http://dx.doi.org/10.1016/j.geomorph.2013.02.024.

Lewin, J., Gibbard, P.L., 2010. Quaternary river terraces in England: forms, sediments and processes. Geomorphology 120, 293–311. http://dx.doi.org/10.1016/j.geomorph.2010.04.002.

Lewin, J., Macklin, M.G., 2003. Preservation potential for Late Quaternary river alluvium. J. Quat. Sci. 18, 107–120. http://dx.doi.org/10.1002/jqs.738.

Lunt, I.A., Bridge, J.S., 2004. Evolution and deposits of a gravelly braid bar, Sagavanirktok River Alaska. Sedimentology 51, 415–432. http://dx.doi.org/10.1111/j.1365-3091.2004.00628.x.

Lunt, I.A., Sambrook-Smith, G.H., Best, J.L., Ashworth, P., Lane, S.N., Simpson, C.J., 2013. Deposits of the sandy braided South Saskatchewan River: implications for the use of modern analogues in reconstructing channel dimensions in reservoir characterization. Bull. Am. Assoc. Pet. Geol. 97, 553–576. http://dx.doi.org/10.1306/09251211152.

Macklin, M.G., Jones, A.F., Lewin, J., 2010. River response to rapid Holocene environmental change: evidence and explanation in British catchments. Quat. Sci. Rev. 29, 1555–1576. http://dx.doi.org/10.1016/j.quascirev.2009.06010.

Makaske, B., Smith, D.G., Berendson, H.J.A., de Boer, A.G., Nielen-Kiezebrink, M.F., Locking, T., 2009. Hydraulic and sedimentary processes causing anastomosing morphology of the upper Columbia River, British Colombia, Canada. Geomorphology 111, 194–205.

Makaske, B., Maathuis, B.H.P., Padovani, C.R., Stolker, C., Mosselman, E., Jongman, R.H.G., 2012. Upstream and downstream controls of recent avulsions on the Taquari megafan, Pantanal, south-western Brazil. Earth Surf. Process. Landf. 37, 1313–1326. http://dx.doi.org/10.1002/esp.3278.

Marchetti, Z.Y., Latrubesse, E.M., Pereira, M.S., Ramonell, C.G., 2013. Vegetation and its relationship with geomorphologic units in the Parana River floodplain, Argentina. J. S. Am. Earth Sci. http://dx.doi.org/10.1016/j.jsames.2013.03.010.

Marriott, S.B., 1996. Analysis and modelling of overbank deposits. In: Anderson, M.G., Walling, M.G., Bates, P.D. (Eds.), Floodplain Processes. Wley, Chichester, pp. 63–93.

Marriott, S.B., Alexander, J., 1999. Introduction. In: Marriott, S.B., Alexander, J. (Eds.), Floodplains: interdisciplinary approaches. Geol. Soc. Lond. Spec. Publ., 163, pp. 1–13.

McCabe, P.J., 1984. Depositional environments of coal and coal-bearing strata. In: Rahmani, R.A., Flores, R.M. (Eds.), Depositional Environments of Coal and Coal Bearing Strata. International Association of Sedimentologists, Special Publication, 7, pp. 13–42.

McCabe, P.J., Parrish, J.T., 1992. Tectonic and climatic controls on the distribution and quality of Cretaceous coals. In: McCabe, P.J., Parrish, J.T. (Eds.), Controls on the Distribution and Quality of Cretaceous Coal. Geological Society of America, Special Paper, 267, pp. 1–15.

Mertes, L.A.K., 1997. Documentation and significance of the perirheic zone on inundated floodplains. Water Resour. Res. 33, 1749–1762.

Mertes, L.A.K., Dunne, T., 2007. The effects of tectonics, climatic history, and sea-level history on the form and behaviour of the modern Amazon River. In: Gupta, A. (Ed.), Large Rivers. Wiley, Chichester, pp. 115–144.

Mertes, L.A.K., Dunne, T., Martinelli, L.A., 1996. Channel-floodplain geomorphology along the Solimões–Amazon River, Brazil. Geol. Soc. Am. Bull. 108, 1089–1107.

Miall, A.D., 2006. Reconstructing the architecture and sequence stratigraphy of the preserved fluvial record as a tool for reservoir development: a reality check. Am. Assoc. Geol. Bull. 90, 989–1002.

Moody, J.A., Pizzuto, J.E., Meade, R.H., 1999. Ontogeny of a flood plain. Geol. Soc. Am. Bull. 111, 291–303.

Moreton, D., Carter, J., Peakcock, M., Rogola, B., 2013. Athabasca Oil Sands: Application of Integrated Technology in the Identification of Commercial Thermal and Mining Opportunities. International Petroleum Technology Conference, Paper 16755, Beijing, China. 13.

Motta, D., Abad, J.D., Langendoen, E.J., Garcia, M.H., 2012. A simplified 2D model for meander migration with physically-based bank evolution. Geomorphology 163–164, 10–25. http://dx.doi.org/10.1016/j.geomorph.2011.06.036.

Musial, G., Reynaud, J.-Y., Gingras, M.K., Féniès, H., Labourdette, R., Parize, O., 2012. Subsurface and outcrop characterization of large tidally influenced point bars of the Cretaceous McMurray Formation (Alberta, Canada). Sediment. Geol. 279, 156–172. http://dx.doi.org/10.1016/j.sedgeo.2011.04.020.

Nanson, G.C., 1980. Point bar and floodplain formation on the meandering Beatton River, northeastern British Columbia, Canada. Sedimentology 27, 3–29.

Nanson, G.C., 1981. New evidence of scroll-bar formation on the Beatton River. Sedimentology 28, 889–891.

Nanson, G.C., Croke, J.C., 1992. A genetic classification of floodplains. Geomorphology 4, 459–486.

Nanson, G.C., Page, K., 1983. Lateral accretion of fine-grained concave benches on meandering rivers. In: Collinson, J.D., Lewin, J. (Eds.), Modern and Ancient Fluvial Systems. International Association of Sedimentologists Special Publication, 6, pp. 133–143.

Nicholas, A.P., 2013. Morphological diversity of the world's largest rivers. Geology.GSA. http://dx.doi.org/10.1130/G34016.1.

Nicholas, A.P., Sandbach, S.D., Ashworth, P.J., Amsler, M.L., Best, J.L., Hardy, R.J., Lane, S.N., Orfeo, O., Parsons, D.R., Reesink, A.J.H., Sambrook Smith, G.H., Szupiany, R.N., 2012. Modelling hydrodynamics in the Rio Paraná, Argentina: an evaluation and intercomparison of reduced-complexity and physics based models applied to a large sand-bed river. Geomorphology 169–170, 192–211. http://dx.doi.org/10.1016/j.geomorph.2012.05.014.

Nicholas, A.P., Ashworth, P.J., Sambrook Smith, G.H., Sandbach, S.D., 2013. Numerical simulation of bar and island morphodynamics in anabranching megarivers. Journal of Geophysical Research Earth Surface 118. http://dx.doi.org/10.1002/jgrf.20132.

Nilsson, C., Reidy, C.A., Dynesius, M., Ravenga, C., 2005. Fragmentation and flow regulation of the world's large river systems. Science 308, 405–408. http://dx.doi.org/10.1126/science.1107887.

Page, K.J., Nanson, G.C., Frazier, P.S., 2003. Floodplain formation and sediment stratigraphy resulting from oblique accretion on the Murrumbidgee River, Australia. J. Sed. Res. 73, 5–14.

Paira, A.R., Drago, E.C., 2006. Genetical, morphological and evolutional relationships of the floodplain lakes in the Middle Paraná River hydrosystem. Z. Geomorphol. 145, 207–228.

Paira, A.R., Drago, E.C., 2007. Origin, evolution, and types of floodplain water bodies. In: Iriondo, M.H., Paggi, J.C., Parma, M.J. (Eds.), The Middle Paraná River. Springer, Verlag: Berlin, pp. 53–81.

Panin, A.V., Sidorchuk, A.Yu., Chernov, A.V., 1999. Historical background to floodplain morphology: examples from the East European Plain. In: Marriott, S.B., Alexander, J. (Eds.), Floodplains: Interdisciplinary Approaches. Geological Society of London Special Publication, 163, pp. 217–229.

Paola, C., Borgman, L., 1991. Reconstructing random topography from preserved stratification. Sedimentology 38, 553–565.

Parker, G., Johannesson, H., 1989. Observations on several recent theories of resonance and overdeepening in meandering channels. In: Ikeda, S., Parker, G. (Eds.), River Meandering. Water Resource Monograph Series, 12. AGU, Washington, D.C., pp. 379–415.

Parker, G., Shimizu, Y., Wilkerson, G.V., Eke, E.C., Abad, J.D., Lauer, J.W., Paola, C., Dietrich, W.E., Voller, V.R., 2011. A new framework for modeling the migration of meandering rivers. Earth Surf. Process. Landf. 36, 70–86. http://dx.doi.org/10.1002/esp.2113.

Parker, N.O., Sambrook Smith, G.H., Ashworth, P.J., Best, J.L., Lane, S.N., Lunt, I.A., Simpson, C.J., Thomas, R.E., 2013. Quantification of the relation between surface morphodynamics and subsurface sedimentological product in sandy braided rivers. Sedimentology 60, 820–839. http://dx.doi.org/10.1111/j.1365-3091.2012.01364.x.

Peakall, J., Ashworth, P.J., Best, J.L., 2007. Meander-bend evolution, alluvial architecture, and the role of cohesion in sinuous river channels: a flume study. J. Sediment. Res. 77, 197–212. http://dx.doi.org/10.2110/jsr.2007.017.

Pizzuto, J.E., 1987. Sediment diffusion during overbank flows. Sedimentology 34, 301–317.

Ramonell, C.G., Amsler, M.L., Toniolo, H.A., 2002. Shifting modes of the Paraná River thalweg in its middle-lower reaches. Z. Geomorphol. Suppl. 129, 129–142.

Reesink, A.J.H., Ashworth, P.J., Sambrook Smith, G.H., Best, J.L., Parsons, D.R., Amsler, M.L., Hardy, R.J., Lane, S.N., Nicholas, A.P., Orfeo, O., Sandbach, S.D., Simpson, C.J., Szupiany, R.N., 2013. Scales and causes of heterogeneity in bars in a large multi-channel river: Rio Paraná. Argentina. Sedimentology. http://dx.doi.org/10.1111/sed.12092.

Reinfelds, I., Nanson, G., 1993. Formation of braided river floodplains, Waimakariri River, New Zealand. Sedimentology 40, 1113–1127.

Rice, S.P., Church, M., 2010. Grain-size sorting within river bars in relation to downstream fining along a wandering channel. Sedimentology 57, 232–251. http://dx.doi.org/10.1111/j.1365-3091.2009.01108.x.

Rice, S.P., Roy, A.G., Rhoads, B. (Eds.), 2008. River Confluences, Tributaries and the Fluvial Network. Wiley, Chichester.

Rice, S.P., Church, M., Wooldridge, C.L., Hicken, E.J., 2009. Morphology and evolution of bars in a wandering gravel-bed river: lower Fraser river, British Columbia, Canada. Sedimentology 56, 709–736. http://dx.doi.org/10.1111/j.1365-3091.2008.00994.x.

Richey, J.E., Nobre, C., Deser, C., 1989. Amazon river discharge and climate variability: 1903–1985. Science 246, 101–103.

Rowland, J.C., Lepper, K., Dietrich, W.E., Wilson, C.J., Sheldon, R., 2005. Tie channel sedimentation rates, owbow formation age and channel migration rate from optically stimulated luminescence (OSL) analysis of floodplain deposits. Earth Surf. Process. Landf. 30, 1161–1179. http://dx.doi.org/10.1002/esp.1268.

Roy, N.G., Sinha, R., Gibling, M.R., 2011. Aggradation, incision and interfluve flooding in the Ganga valley over the past 100,000 years: testing the influence of monsoonal precipitation. Palaeogeogr. Palaeoclimatol. Palaeoecol. http://dx.doi.org/10.1016/j.palaeo.2011.08.012.

Rozo, M.G., Nogueira, A.C.R., Truckenbrodt, W., 2012. The anastomosing pattern and the extensively distributed scroll bars in the middle Amazon River. Earth Surf. Process. Landf. 14, 1471–1488. http://dx.doi.org/10.1002/esp.3249.

Sadler, P.M., 1981. Sediment accumulation rates and the completeness of stratigraphic sections. J. Geol. 89, 569–584.

Sambrook Smith, G.H., Best, J.L., Bristow, C.S., Petts, G.E. (Eds.), 2006. Braided rivers: process, deposits, ecology and management. International Association of Sedimentologists Special Publication, 36. Blackwell, Oxford.

Sambrook Smith, G.H., Best, J.L., Ashworth, P.J., Lane, S.N., Parker, N.O., Lunt, I.A., Thomas, R.E., Simpson, C.J., 2010. Can we distinguish flood frequency and magnitude in the sedimentological record of rivers? Geology 38, 579–582. http://dx.doi.org/10.1130/G30861.1.

Santos, M.L., Stevaux, J.C., 2000. Facies and architectural analysis of channel sandy macroforms in the upper Paraná River. Quaternary International 72, 87–94.

Sidorchuck, A.Y., Panin, A.V., Borisova, O.K., 2009. Morphology of river channels and surface runoff in the Volga River basin (East European Plain) during the Late Glacial period. Geomorphology 113, 137–157.

Sidorchuk, A., 2003. Floodplain sedimentation: inherited memories. Glob. Planet. Chang. 39, 13–29.

Sippel, S.J., Hamilton, S.K., Melack, J.M., 1992. Inundation area and morphometry of lakes on the Amazon River floodplain. Braz. Arch. Hydrobiol. 123, 385–400.

Slingerland, R., Smith, N.D., 2004. River avulsions and their deposits. Ann. Rev. Earth Planet. Sci. 32, 257–285.

Smith, D.G., 1986. Anastomosing river deposits, sedimentation rates and basin subsidence, Magdalena River, Northwestern Colombia, South America. Sediment. Geol. 46, 177–196.

Smith, D.G., Hubbard, S.M., Leckie, D.A., Fustic, M., 2009. Counter point-bar deposits: lithofacies and reservoir significance in the meandering modern Peace River and ancient McMurrary Formation, Alberta, Canada. Sedimentology 56, 655–669. http://dx.doi.org/10.1111/j.1365-3091.2009.01050.x.

Smith, M.W., Macklin, M.G., Thomas, C.J., 2013. Hydrological and geomorphological controls of malaria transmission. Earth Sci. Rev. 116, 109–127. http://dx.doi.org/10.1016/j.earscirev.2012.11.004.

Snedden, J.W., 2013. Channel-body basal scours: observations from 3D seismic and importance for subsurface reservoir connectivity. Mar. Pet. Sci. 39, 150–163. http://dx.doi.org/10.1016/j.marpetgeo.2012.08.013.

Steiger, J., Gurnell, A.M., Petts, G.E., 2001. Sediment deposition along the channel margins of a reach of the Middle River Severn, UK. Regul. Rivers Res. Manag. 17, 443–460.

Stevaux, J.C., Corradini, F.A., Aquino, S., 2013. Connectivity processes and riparian vegetation of the upper Paraná River, Brazil. J. S. Am. Earth Sci. http://dx.doi.org/10.1016/j.jsames.2011.12.007.

Straub, K.M., Paola, C., Mohrig, D., Wolinsky, M.A., George, T., 2009. Compensational stacking of channelized sedimentary deposits. J. Sediment. Res. 79, 673–688. http://dx.doi.org/10.2110/jsr.2009.070.

Strauss, D., Sadler, P.M., 1989. Stochastic-models for the completeness of stratigraphic sections. Math. Geol. 21 (1), 37–59. http://dx.doi.org/10.1007/BF00897239.

Sun, T., Meakin, P., Jossang, T., 1996. A simulation model for meandering rivers. Water Resour. Res. 32, 2937–2954.

Syvitski, J.P.M., Vorosmarty, C.J., Kettner, A.J., Green, P., 2005. Impact of humans on the flux of terrestrial sediment to the global coastal ocean. Science 308, 376–380.

Syvitski, J.P.M., Overeem, I., Brackenridge, G.R., Hannon, M., 2012. Floods, floodplains, delta plains — a satellite imaging approach. Sediment. Geol. 267–268, 1–14. http://dx.doi.org/10.1016/j.sedgeo.2012.05.014.

Thomas, L., 2013. Coal Geology (2nd edition). Wiley-Blackwell, Chichester.

Thorne, C.R., Russell, A.P.G., Alam, M.K., 1993. Planform pattern and channel evolution of the Brahmaputra River, Bangladesh. In: Best, J.L., Bristow, C.S. (Eds.), Braided rivers: Special Publication of the Geological Society of London, 75, pp. 257–276.

Tipper, J.C., 1983. Rates of sedimentation, and stratigraphical completeness. Nature 302 (5910), 696–698. http://dx.doi.org/10.1038/302696a0.

Tockner, K., Stanford, J.A., 2002. Riverine flood plains: present state and future trends. Environ. Conserv. 29, 308–330. http://dx.doi.org/10.1017/S037689290200022X.

Toonen, W.H.J., Kleinhans, M.G., Cohen, K.M., 2012. Sedimentary architecture of abandoned channel fills. Earth Surf. Process. Landf. 37, 459–472. http://dx.doi.org/10.1002/esp.3189.

Tooth, S., 2013. Dryland fluvial environments: assessing distinctiveness and diversity from a global perspective. In: Shroder, J.F. (Ed.), Treatise on Geomorphology, vol. 9. Academic Press, San Diego, CA, pp. 612–644.

Tooth, S., Nanson, G.C., 2011. Distinctiveness and diversity of arid zone river systems, In: Thomas, D.S.G. (Ed.), Arid Zone Geomorphology: Process form and Change in Drylands, 3rd edition. Wiley, Chichester, pp. 269–300.

Tooth, S., Rodnight, H., Duller, G.A.T., McCarthy, T.S., Marren, P.M., Brandt, D., 2007. Chronology and controls of avulsion along a mixed bedrock-alluvial river. Geol. Soc. Am. Bull. 119, 452–461.

Törnqvist, T.E., Bridge, J.S., 2002. Spatial variation of overbank aggradation rate and its influence on avulsion frequency. Sedimentology 49, 891–905.

Trigg, M.A., Bates, P.D., Wilson, M.D., Schumann, G., Baugh, C., 2012. Floodplain channel morphology and networks of the middle Amazon River. Water Resour. Res. 48, W10504. http://dx.doi.org/10.1029/2012WR01188.

Trigg, M.A., Michaelides, K., Neal, J.C., Bates, P.D., 2013. Surface water connectivity dynamics of a large scale extreme flood. J. Hydrol. http://dx.doi.org/10.1016/j.jhydrol.2013.09.035.

Valente, C.R., Latrubesse, E.M., 2011. Fluvial archive of peculiar avulsive fluvial patterns in the largest Quaternary intracratonic basin of tropical South America: the Bananal Basin Central-Brazil. Palaeogeogr. Palaeoclimatol. Palaeoecol. http://dx.doi.org/10.1016/j.palaeo.2011.1.10.002.

Valente, C.R., Latrubesse, E.M., Ferreira, L.G., 2013. Relationships among vegetation, geomorphology and hydrology in the Bananal Island tropical wetlands, Araguaia River basin, Central Brazil. J. S. Am. Earth Sci. http://dx.doi.org/10.1016/j.jsames.2012.12.003.

Van De Wiel, M.J., Coulthard, T.J., 2010. Self-organized criticality in river basins: challenging sedimentary records of environmental change. Geology 38, 87–90. http://dx.doi.org/10.1130/G30490.1.

van de Wolfshaar, K.E., Middlekoop, H., Addink, E., Winter, H.V., Nagelkerke, L.A.J., 2011. Linking flow regime, floodplain lake connectivity and fish catch in a large river-floodplain system, the Volga-Akhtuba floodplain (Russian Federation). Ecosystems 14, 920–934.

Van Dijk, W.M., van der Lageweg, W.I., Kleinhans, M.G., 2012. Experimental meandering river with chute cutoffs. J. Geophys. Res. 117, F03023. http://dx.doi.org/10.1029/2011FJ002314.

Vietz, G.J., Rutherfurd, I.D., Stewardson, M.J., Finlayson, B.L., 2011. Hydrodynamics and sedimentology of concave benches in a lowland river. Geomorphology 147 (8), 86–101. http://dx.doi.org/10.1016/j.geomorph.2011.07.033.

Walling, D.E., 2008. The changing sediment loads of the world's rivers. Sediment Dynamics in Changing Environments, 325. IAHS Publ., pp. 323–328.

Walling, D.E., Moorhead, P.W., 1989. The particle size characteristics of fluvial suspended sediment: an overview. Hydrobiologia 176 (7), 125–149.

Walling, D.E., He, Q., Nicholas, A.P., 1996. Floodplains as suspended sediment sinks. In: Anderson, M.G., Walling, M.G., Bates, P.D. (Eds.), Floodplain Processes. Wiley, Chichester, pp. 399–440.

Wang, Y., Straub, K.M., Hajek, E.A., 2011a. Scale-dependent compensational stacking: an estimate of autogenic time scales in channelized sedimentary deposits. Geology 39 (9), 811–814. http://dx.doi.org/10.1130/G32068.1.

Wang, H., Saito, Y., Zhang, Y., Naishuang, B., Sun, X., Yang, Z., 2011b. Recent changes of sediment flux to the western Pacific Ocean from major rivers in East and Southeast Asia. Earth Sci, Rev http://dx.doi.org/10.1016/j.earscirev.2011.06.003.

Werritty, A., Paine, J.L., Macdonald, N., Rowan, J.S., McEwen, L.J., 2006. Use of multi-proxy records to improve estimates of flood risk: lower River Tay, Scotland. Catena 66, 107–119.

Westaway, R.M., Lane, S.N., Hicks, D.M., 2003. Remote survey of large-scale braided rivers using digital photogrammetry and image analysis. Int. J. Remote Sens. 24, 795–816.

Zinger, J.A., Rhoads, B.L., Best, J.L., 2011. Extreme sediment pulses generated by bend cutoffs along a large meandering river. Nat. Geosci. 4, 675–678. http://dx.doi.org/10.1038/NGEO1260.